THE UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR Georges Roman Dagenais

TITLE OF THESIS The oxygen isotope geochemistry of granitoid rocks from the southern and central Yukon

DEGREE FOR WHICH THESIS WAS PRESENTED Master of Science

YEAR THIS DEGREE GRANTED Spring, 1984

THE UNIVERSITY OF ALBERTA

The oxygen isotope geochemistry of granitoid rocks from the southern and central Yukon

by

© Georges Roman Dagenais

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE

OF Master of Science

Geology

EDMONTON, ALBERTA

Spring, 1984

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled The oxygen isotope geochemistry of granitoid rocks from the southern and central Yukon submitted by Georges Roman Dagenais in partial fulfilment of the requirements for the degree of Master of Science.

Dedication

Cette thèse est dédiée à mes parents,
en reconnaissance de leur encouragement
et de leur aide au cours de mes études.

ABSTRACT

Oxygen isotope ratios of 80 granitoid samples from the southern and central Yukon were determined along with 39 country rocks. The granites are subdivided into two very distinct primary $\delta^{18}O$ groups; a) a primitive set of values ranging from +6.0 to 8.5‰ and b) a high $\delta^{18}O$ group averaging +10‰ or more. The first group tends to crop out in the Mesozoic, arc-related terranes newly accreted to the ancient western margin of the North American plate. The second group is found in old cratonic settings. The final $\delta^{18}O$ composition of the granitoid rocks of the study area is partially controlled by the isotopic composition of the wall rocks.

Alteration patterns emphasize the differences in the above groups: the lower ^{18}O set is depleted by isotopically light meteoric waters (-16‰) during hydrothermal cycling while the high ^{18}O rocks of the eastern terranes preserve their original $\delta^{18}O$ despite alteration. In the Atlin and Omineca Belts, groundwaters must have interacted at low (50-150°C) temperatures with high ^{18}O metasedimentary rocks that increased their $\delta^{18}O$ to values of 0 to +6‰ or more.

Two main hydrothermal periods were identified, one in the Middle Cretaceous era (110 to 90 Ma), and the other in Tertiary times (65 to 50 Ma). K-Ar ages of plutons have been reset to the above hydrothermal periods by the large scale heating present during climactic plutonic activity.

Comparison of Sr isotopic and ^{18}O data delineates three groups of rocks with different petrogenetic histories: (1) a primitive ^{18}O and low initial $^{87}Sr/^{86}Sr$ group (< 0.706) from a deep-seated, arc-related source having undergone little or no upper crustal interaction, (2) a high ^{18}O and radiogenic Sr group (>0.710) derived from anatexis or interaction with isotopically enriched cratonic sediments and (3) A low-^{18}O and moderate (>0.706) initial $^{87}Sr/^{86}Sr$ group of calc-alkaline intrusives in the Yukon Crystalline Terrane, that may have been generated by subduction processes involving marine sediments, or alternatively, were produced at lower crustal depths and interacted with a radiogenic Sr source, during their ascent in the upper crust.

Acknowledgements

Field work on this project was supported by the Boreal Institute for Northern Studies grant-in-aid program. Laboratory expenses were covered by NSERC grant # A9165. I wish to thank D.J. Tempelman-Kluit and J. Morin and the D.I.A.N.D. staff in Whitehorse for their help during field work in the Yukon and also R. Armstrong and B. Ballantyne for providing some of the samples for this study. Special thanks are also given to Mrs. E. Toth, for her help in the laboratory, and to Dr. K. Muehlenbachs for his guidance and help in the course of this study.

Table of Contents

List of Tables

List of Figures

Abbreviations

For the purpose of brevity, the following terms will be abbreviated in the manner shown, where deemed necessary;

Yukon Crystalline Terrane.................................YCT

Whitehorse Trough...WT

Omineca Crystalline Belt..................................OCB

Coast Plutonic Complex....................................CPC

Quartz..Qtz

Feldspar..Fp

K-Feldspar..Kfp

Plagioclase...Pl

Biotite...Bio

Hornblende..Hbl

Magnetite...Mag

Sericite..Ser

Chlorite..Chl

Epidote...Epi

Pyrite..Py

Calcite...Ca

Argon...Ar

Potassium...K

Rubidium..Rb

Strontium...Sr

I. INTRODUCTION

A. The goal

The measurement of oxygen isotope ratios in plutonic rocks has become widespread in recent years. The systematics of mineral isotope partitioning permit the identification of pre- and post-crystallisation processes affecting cooling plutons. These processes include weathering, hydrothermal alteration, assimilation of country rocks, pluton - wall rock exchange, among others.

The main goal of this study was to measure the relative importance of post-solidus ^{18}O exchange on the final isotopic composition of granites of the northern Cordillera. An attempt was made to estimate the physical and chemical conditions present during periods of hydrothermal activity. Also primary ^{18}O concentrations in the plutons were compared to Rb-Sr and K-Ar data from previous studies to gain additional insight into the petrogenetic history of the rocks.

Data gathering

Some fifty sample splits of granitoid rocks from previous radioisotopic studies (Bultman, 1979, Morrison *et al.*, 1979, Godwin *et al.*, 1980 and R. Armstrong, unpublished data) were obtained for this study. For rocks from the Yukon Crystalline Terrane, splits were no longer available and resampling of original locations (LeCouteur &

Tempelman-Kluit, 1976) was undertaken with the help of D.J. Tempelman-Kluit to permit comparison of our $\delta^{18}O$ results with Sr isotopes. Supplemental sampling over the entire study area was done to obtain a better coverage of each of the geological provinces. Samples of about 2-3 kg. were generally taken well away from contacts except where it was useful to study country rock-batholith interaction.

In all cases the freshest available specimens were obtained from the outcrop and therefore, it is thought that the results outlined below are characteristic of the main body of the batholiths in the study area. Surface weathering was removed in the field and in the laboratory until a homogeneous ("fresh") sample resulted. The rocks and powders obtained from other workers, were collected initially for radioisotopic measurements by the original authors, and the same general sampling procedures were used in these rocks as outlined above (R. Armstrong, personal communication).

Location and access

The main study area is bound respectively to the west and east by the Denali-Shakwak and Tintina faults located in the southern and central parts of the Yukon Territory (figure 1-1). The northernmost extension of sampling includes the Coffee Creek quartz monzonite (63°N), while rocks in the Atlin Lake area of northern British Columbia (59°N) define the southern limit of study. Two samples from the Selwyn Basin granitic stocks, beyond the Tintina Trench

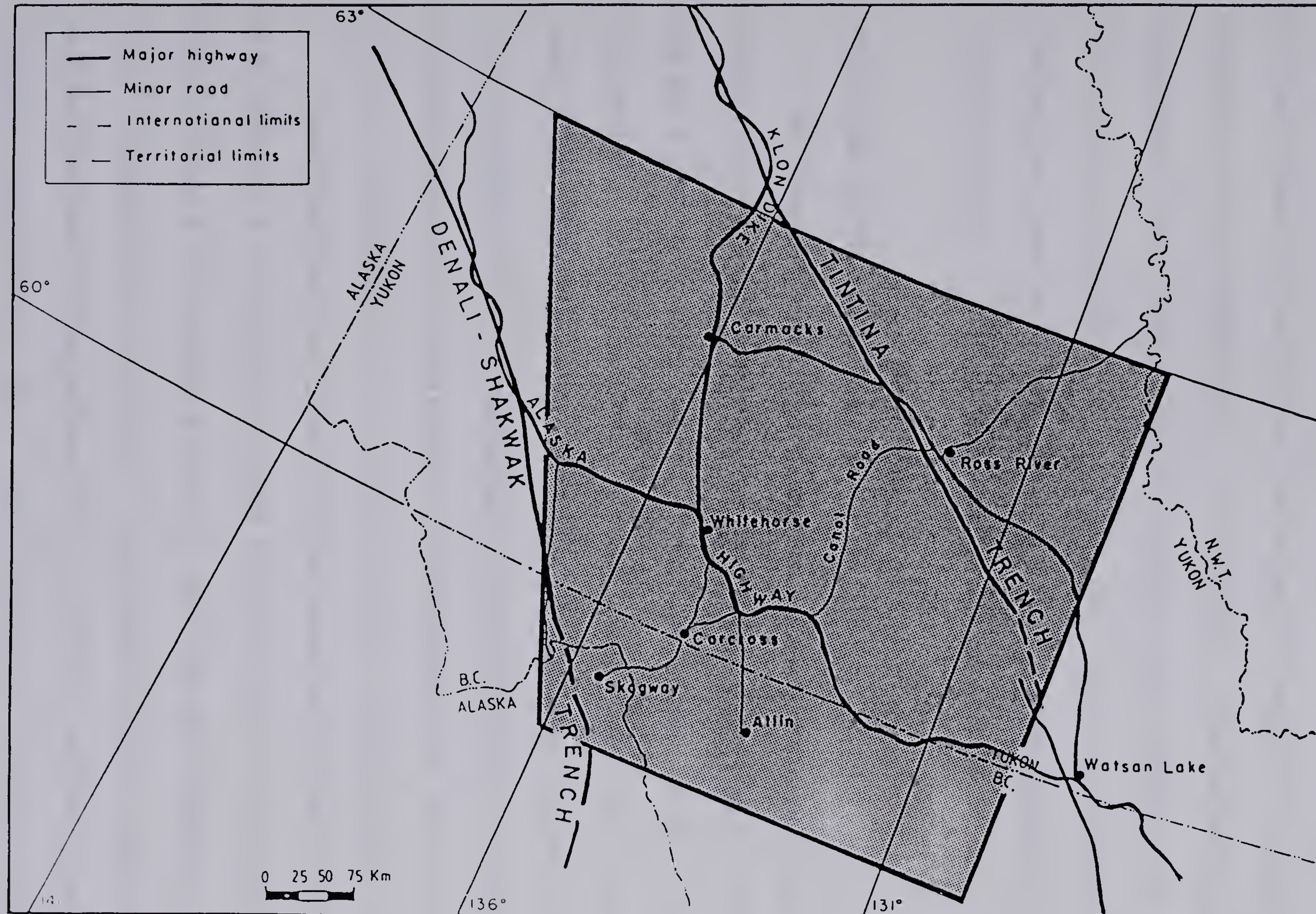


Figure 1-1 Location map of study area. The shaded area outlines the region in which the samples of this study crop out. Access routes and major population centers are also shown.

were also analysed (43-D & KTP).

Access to the area is obtained by the Alaska, Klondike and Faro highways along with other subsidiary roads. Access to the Yukon Crystalline Terrane is mostly limited to air travel and backpacking. A helicopter chartered out of Carmacks, Y.T. was used to collect the samples of this area. The high latitude and altitude of the region meant overburden was minimal, which facilitated sampling considerably.

Previous work in the area

Extensive geological reconnaissance mapping by the Geological Survey of Canada and the Department of Indian Affairs and Northern Development on a scale of 1:250 000 † has been accomplished since the early part of the century. Gabrielse and Wheeler (1961) provided the first tectonic, stratigraphic and structural interpretation of the Whitehorse area. Subsequent syntheses were done by Monger (1975), Tempelman-Kluit (1976, 1979) and Bultman (1979).

Geochemical studies in the area are mostly centered on ore deposits and mineralised claims with few exceptions. The most notable of these are K-Ar and Rb-Sr studies of the Yukon Crystalline Terrane (Tempelman-Kluit & Wanless, 1975 and LeCouteur & Tempelman-Kluit, 1976) and of the Whitehorse Trough (Morrison *et al.*, 1979, Bultman, 1979). Previous

----

† See Yukon geology and exploration 1979-80, D.I.A.N.D. publication # QS 8278-000-E1-A1, for a complete listing of geological maps and memoirs available.

oxygen isotope work within the area is represented by the Skagway-Yakutat Bay traverses (Magaritz & Taylor, 1976a) and some data in G. Morrison's thesis (1981). Ongoing isotope work is concentrated on tin-tungsten deposits in the Selwyn Fold Belt (Longstaffe *et al.*, 1982, Anderson, 1983a, 1983b, Bowman & Covent, 1983).

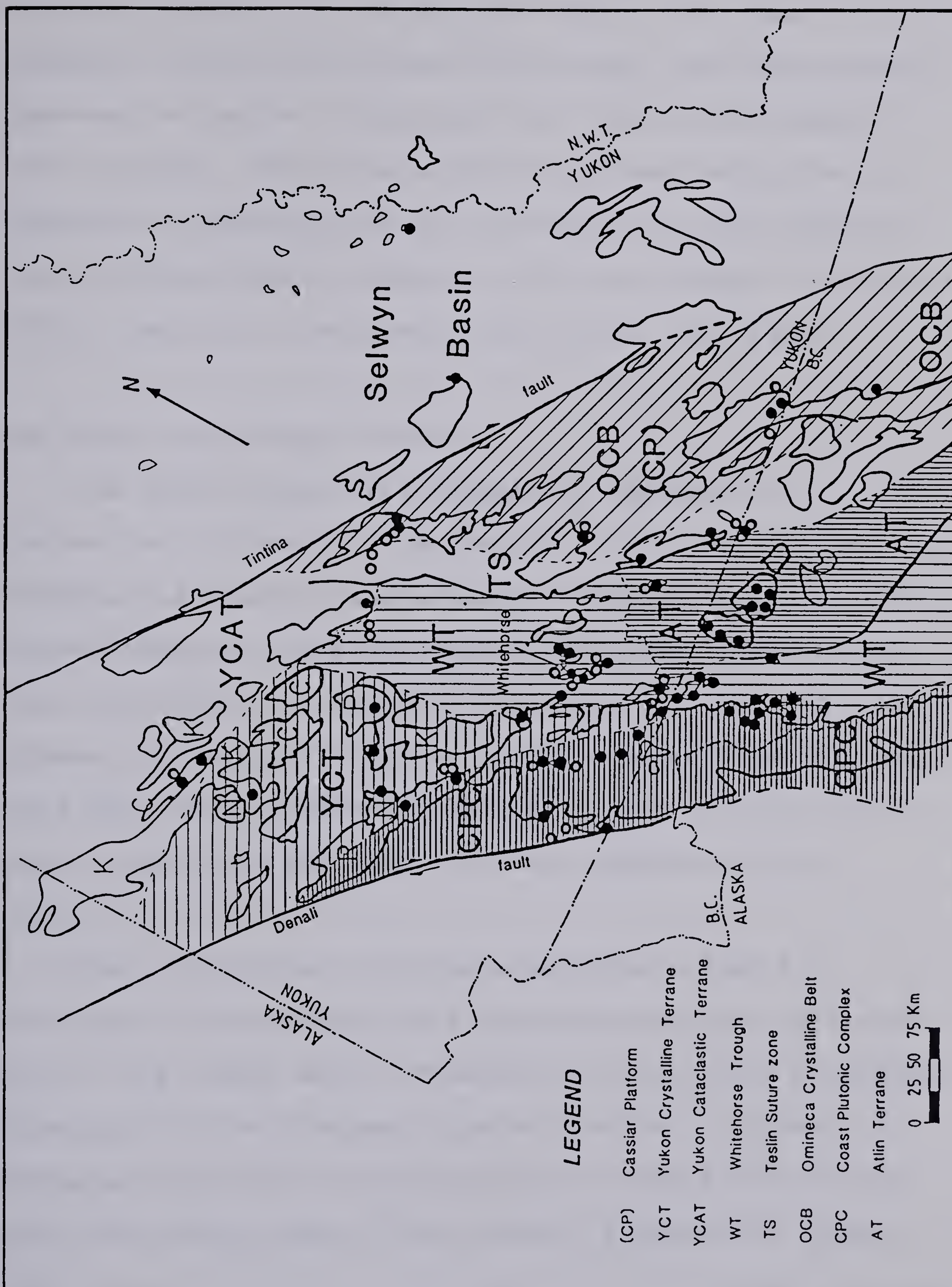
B. Regional geology

The Canadian cordillera is subdivided into five tectonic provinces including, from west to east, (1) the Insular Fold Belt, (2) the Coast Plutonic Complex, (3) the Intermontane Belt, (4) the Omineca Crystalline Belt, and (5) the Eastern Fold Belt, four of which (2-5) were sampled during the course of this study (figure 1-2).

Coast Plutonic Complex

The Coast Plutonic Complex is composed of medium to high grade metamorphic rocks and migmatitic complexes (Hutchison, 1970) of Upper Paleozoic to Upper Tertiary age with some remnants of Lower Paleozoic and Precambrian supracrustal rocks. The belt is intruded by a large number of calc-alkaline intrusives of mainly Lower Cretaceous to Tertiary age (Roddick & Hutchison, 1974) that coalesced to form an almost continuous batholith from Vancouver B.C. to the Yukon Territory. The western part of the belt shows a variety of structural and textural relationships between igneous and metamorphic rocks that include migmatite

Figure 1-2 Tectonostratigraphic map of study area. The five geological terranes outlined in the text are shown. The YCT, CPC, and WT units are called the western terranes, while the AT, OCB and Selwyn Basin make up the eastern terranes of the study. The YCAT is a subdivision of the YCT; it is comprised of metamorphic rocks of sedimentary (clastic) origin. The TS separates the newly-accreted Mesozoic units from the old North American plate margin. General sample locations are shown; filled dots represent granitoid samples, open circles, country rocks. The batholiths are outlined in ink. The letters found in the YCT identify the respective rock suite; K → Klotassin suite, P → Porphyritic suite, C → Coffee Creek qtz monzonite, and N → Nisling Range alaskite suite. The political boundaries are illustrated in the same way as in figure 1-1.


LEGEND
(CP) Cassiar Platform
YCT Yukon Crystalline Terrane
YCAT Yukon Cataclastic Terrane
WT Whitehorse Trough
TS Teslin Suture zone
OCB Omineca Crystalline Belt
CPC Coast Plutonic Complex
AT Atlin Terrane
0 25 50 75 Km
N
Selwyn Basin
Tintina fault
Denali fault
N.W.T.
YUKON
B.C.
ALASKA
Whitehorse

complexes with gradational contacts to discordant, clearly intrusive plutons. In the eastern edge of the Coast Mountains, plutons are clearly discordant and often contain numerous inclusions of country rocks (Gabrielse & Reesor, 1974, Bultman, 1979). The bulk of the Coast batholith is composed of magnetite-series, so-called "I" type granitoid rocks as described by Ishihara (1977) and Chappell and White (1974), implying a deep seated source for the magmas.

The Yukon Crystalline Terrane

The Yukon Crystalline Terrane is composed of Precambrian and Paleozoic metamorphic rocks such as the Klondike and Kluane Schists (mafic), the Pelly gneiss, the Nasina quartzite, marbles, amphibolites and serpentinites. They are broadly correlated to similar lithologies in the adjacent Omineca Belt and are interpreted as metamorphosed shelf and slope sediments deposited at the ancient western edge of the North American continent (Tempelman-Kluit, 1976).

Four intrusives suites have been identified by structural, stratigraphic and age dating methods (K-Ar and Rb-Sr). The oldest suite (Triassic?) is the large discordant batholiths of the Klotassin quartz diorite. It forms elongated north-west trending plutonic masses that occupy most the central area of the terrane (K symbols in figure 1-2).

The pink quartz monzonite and the porphyritic biotite quartz monzonite suites (Tempelman-Kluit, 1976) are somewhat younger than the Klotassin rocks and show gradational to sharp contacts with it (P symbol in figure 1-2). Both of the above suites are coarse grained, slightly foliated and are thought to represent a Mesozoic arc root called Stikinia after the Upper Triassic Stikine Arch in central British Columbia.

The Mid-Cretaceous rocks (90-100 Ma) of the Coffee Creek quartz monzonite are found in the northern part of the terrane. The suite is compositionally restricted and shows very sharp contacts with all older rocks in the area. It is thought to be the equivalent of Mid-Cretaceous batholiths in the Omineca Belt (Gabrielse & Reesor, 1974). Tertiary intrusive rocks are represented by the Nisling Range alaskite † suite (50-60 Ma). It consists of rounded strongly discordant stocks and batholiths that intrude the southwestern part of the terrane. Porphyritic textures, myrmekitic and micrographic quartz and miarolitic cavities all point to an epizonal nature for these rocks. Associated dyke swarms and explosive felsic volcanics of the Mt-Nansen, Sloko and Skukum groups crop out in the immediate area and also in the Intermontane Belt further south. Finally, the Ruby Range granodiorite, a medium-grained equigranular hornblende and biotite-bearing batholith outcrops at the south-western edge of the Yukon Crystalline Terrane. K-Ar

----

† Alaskite is another name for alkali-feldspar granite

ages in this suite vary from 175 to 49 Ma, probably the result of partial to complete resetting by the younger Nisling alaskite (Tempelman-Kluit & Wanless, 1975). The suite has many affinities with the coast plutons and is thought to be the northern extension of that belt.

The Whitehorse Trough

The Whitehorse Trough is the northern part of the Intermontane Belt and can be described as a northwest trending synclinorium (Bultman, 1979). Lower to Middle Jurassic sedimentary rocks (conglomerates, arkosic and quartzitic sandstones, siltstones and argillites) conformably to uncomformably overly Upper Triassic basalts and andesites, pyroclastic rocks, volcaniclastic rocks and limestone reefoid masses (Wheeler, 1961, Bultman, 1979). The latter rocks, called the Lewes River group, are an island arc assemblage that outcrops at the edges of the Trough. The younger, successor basin-type sedimentary rocks of the Laberge and Tantalus groups, lie in the middle of the area, along the fold axis of the synclinorium.

The Whitehorse plutons intrude into the Triassic Lewes River group and the lower part of the Laberge group in the center of the Trough. They form small to medium sized <200 km^2, elongated, medium- to fine-grained granodiorites and quartz monzonites as well as sub-rounded coarse-grained granitic stocks. Contacts with the country rocks are sharp with numerous inclusions present at the periphery of the

plutons. These intrusive rocks were probably emplaced by stoping of roof rocks (Bultman, 1979).

Eastern Terranes

The Atlin Terrane is a wedge shaped structural block consisting of Late Paleozoic and Mesozoic rocks of oceanic derivation. They include chert, argillite, basalt, alpine-type ultramafic rocks, and limestones, metamorphosed to prehnite-pumpellyite or lower greenschist facies (Monger, 1975). The terrane is considered to be part of an obducted Paleozoic oceanic crust and is a subdivision of the Cache Creek Terrane that lies along the eastern margin of the Intermontane Belt (Monger *et al.*, 1982).

Intrusive rocks crop out as large (>500 km^2) intermediate to acid batholithic masses in the center of the terrane. Smaller granitic stocks intrude mostly volcanic rocks at the margins of the region. Contacts are clearly intrusive in both types of plutons, often accompanied by structural disruptions and thermal metamorphism (Bultman, 1979).

The Teslin suture zone is a narrow band of cataclastically deformed material separating rocks of the Atlin Terrane from those of the Omineca Belt. The suture is thought to divide original Precambrian North American continental crust to the east from the newly accreted terranes to the west.

In the Omineca Belt, miogeoclinal rocks of Precambrian to Paleozoic age and Mesozoic volcanic rocks and pelites are intruded by large Late Jurassic to Middle Cretaceous batholiths. Country rocks are metamorphosed up to high grades (7 Kb of pressure due to burial). The Plutonic rocks clearly show intrusive relationships (Gabrielse & Reesor, 1974) and are compositionally restricted to quartz monzonites.

C. Analytical techniques

Most minerals analysed in this study were purified by the author by standard separation techniques, including, hand picking, heavy liquid flotation and magnetic separation. Quartz samples were purified to better than 95% by reaction with fluoboric acid (48%) or silica saturated fluosilicic acid for 2 or more hours at 200-300°C. All other separates were purified to >90% with most samples containing less than 5% impurities. Oxygen yields during the extraction procedure confirmed the purity of the separates as they were within 5% of optimum levels. No attempt was made to separate chlorite found within altered biotites and amphiboles. Sample WHA 6f is an exception to the norms described above, as its very small size did not permit very efficient separation of quartz from feldspar (chap.3).

Oxygen was extracted from silicates and oxides by reaction with BrF_5 at 630°C as described by Clayton and Mayeda (1963). Carbonates were reacted with 100% H_3PO_4 at

25°C for 6 or more hours (McCrea, 1950). Oxygen from water samples, including the SMOW and SLAP standards were equilibrated with tank CO_2 for one week (Epstein and Mayeda, 1953).

Results are reported in the usual $\delta^{18}O$ notation in permil (°/oo) relative to SMOW (Craig, 1961a) or PDB (Craig, 1957). The pooled residual standard deviation of replicate silicate samples was 0.18°/oo for whole-rock and feldspar analyses, 0.14°/oo for quartz. The standard deviation of 12 NBS-28 results run during the course of this study was 0.05°/oo. Isotopic measurements were performed using CO_2 samples in a Nier-type, double collecting, 90° sector mass spectrometer (Micromass 602D).

The fractionation between two phases, a and b, is defined as $1000 \ln\alpha(a-b) \cong \Delta^{18}O_{a-b} = \delta^{18}Oa - \delta^{18}Ob$, where $\alpha(a-b)$ is the ^{18}O ratio between the two phases, R(a)/R(b).

Standards

Twelve NBS-28 measurements gave an average $\delta^{18}O$ of +9.37°/oo ± 0.05 (SMOW) when compared to an internal laboratory carbonate standard which was calibrated using V-SMOW and SLAP waters and corrected for all instrumental and procedural errors. A fractionation factor of 1.04073 (O'Neil & Epstein, 1966a) was assumed between CO_2 and water at 25°C for the purpose of calibrating the internal standard with respect to the SMOW scale. Recently, new measurements of αCO_2-H_2O from different labs and revised mass

spectrometer corrections have produced an average value of 1.0412 (O'Neil *et al.*, 1975, Friedman & O'Neil, 1977). When comparing our data with studies using the revised α value, $0.47°/_{oo}$ should be added to my results.

By far the most abundant source of ^{18}O data on granites comes from the Caltech labs (H.P. Taylor Jr.). Onuma and others (1972) showed that there was a $0.3°/_{oo}$ positive shift in the Caltech results when compared to the Chicago lab. The NBS-28 quartz results from this study are shifted $\simeq +0.2°/_{oo}$ relative to Chicago (Ito & Clayton, 1983). Therefore, when comparing my results to H.P. Taylor's data, $0.1°/_{oo}$ should be added to the $\delta^{18}O$ values presented below.

II. RESULTS

In this chapter, the studied rocks will be described briefly. The isotope data obtained will be related to and in some cases, explained by the petrographic and geological features of the samples. Radiometric age data will also be quoted, where it may be pertinent to the understanding of the history of the rocks. The purpose of this section is to emphasize both the similarities and the differences in the isotopic signatures of large areas of geology, and to relate them to processes from which they were derived. For a summary of the petrographic features outlined in this section and for the exact locations of the samples, refer to appendices 1 and 2.

A. Whitehorse Trough granitoid rocks

The analytical data of 28 samples of granitoid rocks from the Whitehorse Trough area are given in Table 2-1. The whole rock values average +4.4°/₀₀ with a range of -0.8°/₀₀ to +8.4°/₀₀. The mean is somewhat skewed toward the low range for granites because of multiple sampling of highly altered rocks. When the arithmetic mean of $\delta^{18}O$ ratios for each discrete pluton is used, the average for the region shifts to +5.5°/₀₀. This is still below the +6.0°/₀₀ minimum for primary ^{18}O in granites (Taylor, 1974a) and is a strong indication that a large body of rocks in this area have reequilibrated with a low ^{18}O reservoir. Five of eight batholiths sampled show signs of meteoric hydrothermal

Table 2-1 : $\delta^{18}O$ of Whitehorse Trough granitoid rocks

Sample	Rock Type	Whole Rock	Qtz	Fp	Bio	Hbl	Mag
Mount McIntyre batholith							
WHA 5a	GD	0.8				3.9	
WHA 6a	Grp	4.1	7.7	0.9		4.5	1.9
WHA 6b	Grp	-0.1					
WHA 6c	Grp	-0.7					
WHA 6d	Grp	-0.8					
WHA 6e	Grp	-0.4					
WHA 6f	Grp	0.7	5.1	3.2			-1.5
Whitehorse batholith							
WHB 1	QM	6.0	8.3	6.7	3.1	6.4	-3.0
WHB 2	GD	5.9					
WHB 3	GD	5.1	8.1	5.3			-1.1
WHB 4	GD	6.9	8.0	6.7			
WHB 5	QD	5.3	8.6	5.3	1.7	4.2	-0.5
KTgd-1	GD	5.6	8.6	5.0	4.5		
KTgd-2	GD	5.8	8.4	6.0			
KTgd-4	GD	5.7	8.5	6.1	2.3	4.5	0.5
Cap Mountain granitoids							
WHA 2a	GD	1.9	7.6	2.4	-0.6	2.3	0.6
WHA 2b	GD	2.5					
WHA 2c	GD	2.2	7.4	3.8	-0.5		0.3
WHA 3a	QM	4.7	7.3	3.8	-0.6		0.4
WHA 3b	QM	4.5					
WHA 3c	QM	3.2					
Mount-Lorne							
WHA 7	QM	7.6	8.5	7.5	3.2		
Montana Mountain granite							
WHA 9	QM	8.3	10.1	8.0			
Bee Peak granites							
T75 101-4	QM	7.6	8.8	7.2			
T75 102-2	QM	8.4	10.1	8.5			
T75 310-2	GD	0.7	7.3	-1.6			
T75 413-1	GD	7.2	8.3	7.0			
Lower Tertiary granite							
WHA 4	QM	6.6	8.9	5.7			

All results are in permil (SMOW). Rock abbreviations are : QM = qtz-monzonite, GD = granodiorite, Grp = granophyre.

alteration. A tendency toward higher degrees of exchange has been noted in finer-grained rocks.

The Mt-McIntyre batholith is an elongated 30 km^2 medium grained augite-bearing hornblende-biotite granodiorite. A fine grained granophyric quartz monzonite is located at the approximate geographic center and topographic high of the intrusion. One sample from the granodiorite (WHA 5) and six from the granophyre (WHA6 a to f) were analysed for whole-rock and mineral $^{18}O/^{16}O$ ratios. These samples were collected by previous workers (Morrison *et al.*, 1979) for the purpose of radiometric age determination. All of these seven samples are poor specimens for age dating as they show highly pronounced alteration of feldspars and chloritisation of hornblende and biotite in thin sections. Uralite, epidote and oxides are also abundant. The average whole-rock $\delta^{18}O$ is + 0.5°/₀₀. All major phases have had their $^{18}O/^{16}O$ ratios lowered through interaction with low ^{18}O water. Feldspars are the most markedly affected by this process and in one case (WHA 6a) has a lower $\delta^{18}O$ than coexisting magnetite (fp = +0.9 vs. mag = +1.9).

Eight granite samples (WHB 1 to 5 & KTgd-1, 2, 4) were analysed from the Whitehorse batholith, which is a mid-Cretaceous 100 km^2 coarse to medium-grained biotite-hornblende granodiorite, cropping out west of the territorial capital. It is topographically lower than the Mt-McIntyre granite further west. Numerous inclusions of Lewes River strata are found inside the intrusion near the

contact zone where partial digestion and thermal metamorphism of stoped blocks and surrounding country rocks is apparent. Thin sections indicate most are moderately-to-lightly altered with plagioclase and biotite usually more affected than K-feldspar and hornblende. Most samples show slight oxygen isotopic disequilibrium between coexisting minerals. Two exceptions stand out (WHB 1 = +6.0‰ & WHB 4 = +6.9‰) which have $\Delta^{18}O$ quartz-feldspar of less than 2‰ and $\Delta^{18}O$ quartz-biotite in the range of 5 to 6‰. This suggests an approach to equilibrium. The WHB 4 sample is problematical; pronounced saussuritisation of plagioclase (+6.7‰) is seen in thin section but the $\Delta^{18}O$ quartz-feldspar is equal to 1.3‰, typical of unaltered granites. The presence of ^{18}O-enriched meteoric water is a possible explanation for this apparent contradiction.

Two separate suites were sampled (Morrison *et al.*, 1979) on Cap Mountain to the east of Whitehorse. The three samples from the WHA 3 suite range in composition from a biotite granite porphyry (3c) to a fine-grained leucocratic, pink quartz monzonite (3b) to a coarser-grained pink biotite monzo-granite (3a). Heavy alteration is visible on all phases in the rocks except quartz. Fully chloritised biotite pseudomorphs are present in WHA 3c were feldspar megacrysts have a more turbid appearance than their groundmass counterparts. The whole-rock $\delta^{18}O$ average for this suite is +4.1‰ confirming the petrographic interpretation above.

The second group of samples from Cap mountain (WHA 2a, b, c) is comprised of a more mafic medium- to fine-grained hornblende-biotite quartz diorite to granodiorite. Its whole-rock average $\delta^{18}O$ is approx. +2.2°/oo and this, as in the first group, indicates that isotopic reequilibration has occurred in this suite. In sample WHA 2a, the $\Delta^{18}O$ fractionations of quartz-feldspar (+5.2°/oo), quartz-hornblende (+5.3°/oo) and quartz-biotite (+8.2°/oo) are not primary igneous fractionations. The feldspar ($\delta^{18}O$ = +2.4°/oo), hornblende (+2.3°/oo) and biotite (-0.6°/oo in this sample, could be depleted in ^{18}O by as much as 5.0, 3.0 and 5.0°/oo respectively, from typical values seen in fresh granites (Taylor, 1968).

Only a slight sericitisation is observed on the surface of a few plagioclase grains. Hornblende is relatively fresh with some progressive discoloration from brownish cores (polarised light) to green rims, suggesting some sort of exchange. Biotite is an accessory phase and shows only traces of chlorite within it. Without significant mineral hydration products, the isotopic exchange described above might be explained by a higher than average temperature of alteration (>400°C) in the hydrothermal system. Such temperatures would be beyond the normal stability range of chlorite and clay minerals.

South of Whitehorse, a group of subcircular to elongated plutons intrude Paleozoic greenstones and clastic rocks (Taku group), Jurassic conglomerates (Laberge Group)

and middle Cretaceous mafic volcanic rocks (Hutshi Group). The rocks are coarse- to medium-grained biotite quartz monzonites which often are rimmed by biotite granodiorite. Hornblende is present as an accessory phase while chlorite and epidote are locally prominent (Bultman, 1979).

Six samples collected by previous workers (Morrison *et al.*, 1979, Bultman, 1979) were analysed from these intrusives. Samples WHA 7, a medium grained granodiorite from Mt-Lorne ($\delta^{18}O$ = +7.6°/₀₀), WHA 9, a porphyritic quartz monzonite ($\delta^{18}O$ =+8.3°/₀₀) from Montana mountain and three samples (T75-101-4, -102-2, -413-1) of porphyritic quartz monzonite from the Atlin area ($\delta^{18}O$ +7.6, +8.4 and +7.2°/₀₀ respectively), all show oxygen isotope ratios which lie within of the 'normal' granite range. Thin sections, of two available samples (WHA 7&9) show very little, if any, traces of alteration of mineral phases. One last sample, T75-310-2, a porphyritic granodiorite taken near Tagish lake B.C., was reported (Bultman, 1979) to be significantly more altered than the other three T75 samples. The $\delta^{18}O$ value for this sample (+0.7°/₀₀) confirms this.

A single sample of coarse-grained leucocratic biotite epigranite (WHA 4) of Eocene age was analysed from the Jackson creek area. The rock contains some megacrysts of microcline as well as hornblende in equal amount to biotite ($\simeq$ 3% each). This type of granite crops out mostly in the Coast Plutonic Belt and Yukon Crystalline Terrane and is thought to be related to subvolcanic and volcanic phases of

the Sloko group in northern B.C. and the Skukum group in the Whitehorse area. Locally these rocks are associated with cauldron subsidence complexes (Lambert, 1974) that often are the locus of hydrothermal activity (Criss & Taylor, 1982).

A thin section of WHA 4 shows moderate-to-intense chloritisation of ferromagnesian minerals as well as the presence of oxides and epidote. Feldspars are transformed to white micas and clays with K-feldspar being more affected by the alteration than coexisting highly zoned plagioclase. The whole-rock $\delta^{18}O$ (+6.6°/₀₀) and the $\Delta^{18}O$ quartz-feldspar (+3.2°/₀₀) indicate some interaction with a low ^{18}O fluid, albeit to a lesser extent than the granodioritic intrusions to the east.

B. Yukon Crystalline Terrane batholiths

The Yukon Crystalline terrane plutons were more sparsely sampled than their more southern counterparts in the Whitehorse Trough or the Coast Plutonic Complex. This is mainly due to the relative inaccessibility of the region by road. One to two samples per batholith, centrally located, is the maximum coverage that was permited by limited helicopter sampling (figure 2-1). Care was taken to resample original target locations from a previous radiogenic isotope study (LeCouteur & Tempelman-Kluit, 1976) to allow comparison of oxygen and strontium isotope data. Minerals were separated and whole-rock powders obtained from a total of twelve samples from eight granitic plutons (Table 2-2).

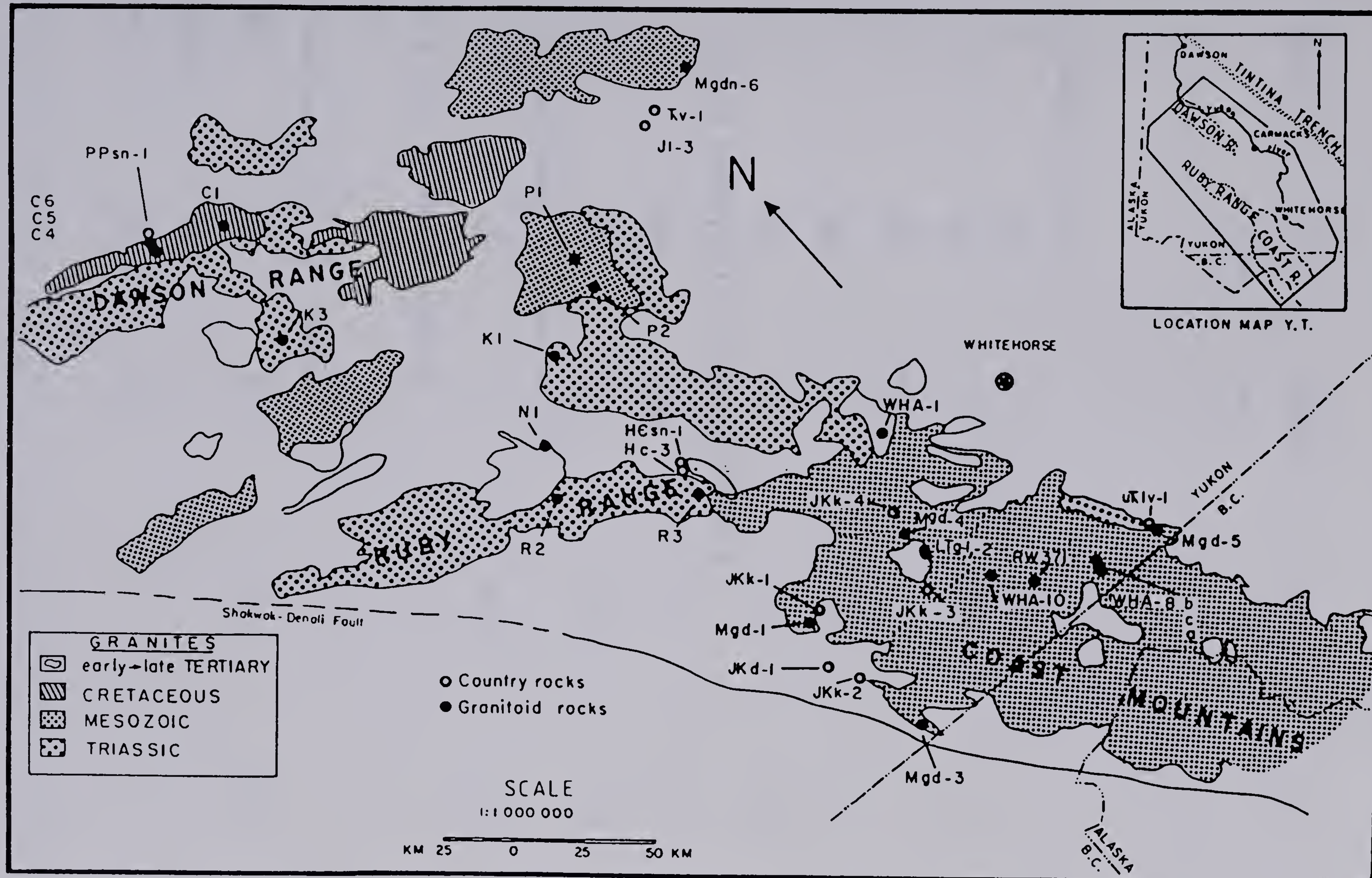


Figure 2-1 Sample locations in the Yukon Crystalline Terrane and in parts of the Coast Range Batholith. The samples are plotted in their approximate position in schematically outlined plutonic masses. Exact positions in degrees and minutes are given in appendices 2 and 4.

Table 2-2 : $\delta^{18}O$ of the Yukon Crystalline Terrane granites

Sample	Rock Type	Whole Rock	Qtz	Fp	Bio
Klotassin suite					
C4	QD	6.6			
C5	QD	-1.0	6.4	-3.7	
C6	QD	7.4			
K1	QD	5.9	8.8		
K3	QD	6.0	8.8	4.8	1.2
Biotite porhyritic qtz-monzonite					
P1	QM	8.0	10.1	8.7	
P2	QM	7.7	8.9	7.4	
Coffee Creek qtz-monzonite					
C1	QM	8.1	9.8	8.3	
Nisling Range alaskite					
N1	Al	9.7			
"Southern Prong" rocks					
WHA 1	Gr	8.5	9.0	7.7	1.9
Mgd-5	GD	-0.2			
WHA 11	GD	6.7			

All results are in permil (SMOW). Al = alaskite, all other abbreviations as before.

Three representative samples of country rock were also analysed but the results will not be presented here; they will be discussed in chapter IV.

The granitoid rocks of this region average $+6.4^{\circ}/_{oo}$ for whole-rock values, which is on the low end of the granitic range (Taylor, 1977). This is mostly due to the statistical predominance of altered rocks in the area, probably caused by multiple phases of intrusive activity, as documented by radiometric ages and geological mapping. The mean isotopic composition of this area is somewhat richer in ^{18}O than the one reported for the Whitehorse Trough ($+5.5^{\circ}/_{oo}$) mostly because of the inclusion of two highly silicic, ^{18}O rich, leucocratic epigranites (N1 and WHA1). This type of granite is not common in the Whitehorse area, therefore no samples of this type of rock were analysed. If the enriched samples above are removed from the mean calculation, the other samples (mostly granodiorites and quartz diorites) from the Yukon Crystalline Terrane give an average $\delta^{18}O$ of $+5.7^{\circ}/_{oo}$; very similar to the one obtained in the Whitehorse area.

No geographic trends were observed in the $\delta^{18}O$ data set for this region, as opposed to the general decrease of Sr isotope initial ratios (LeCouteur & Tempelman-Kluit, 1976) southwestward across the Yukon Crystalline Terrane.

The Klotassin suite (K1, K3 , C4, C5, C6) is comprised of coarse-grained rocks of granodioritic and quartz dioritic composition. The analyses show (Table 2-2) low $\delta^{18}O$ results which are indicative of meteoric hydrothermal alteration.

Samples near the Casino porphyry copper complex (C4, C5, C6), are highly variable in $\delta^{18}O$ along distances of only a few hundred meters. C5, a coarse grained biotite-hornblende quartz diorite ($\delta^{18}O$ =-1.0°/₀₀) has definitely interacted with low ^{18}O meteoric water. The surrounding samples (C4 & C6), have a more ambiguous isotopic signature. Saussuritisation of feldspars and partial chloritisation of biotite as well as the formation of epidote in hornblende (C6), all indicate that significant alteration has occurred in these rocks. Their isotopic compositions are therefore not considered primary (cf. chapter 3).

The K3 sample, from the southern extension of the Klotassin batholith, is a coarse-grained porphyritic hornblende-biotite granodiorite - tonalite. It is moderately saussuritised and has a $\delta^{18}O$ that is quite low (+6.0°/₀₀). Disequilibrium is indicated by the relatively large fractionations between coexisting mineral phases (figure 2-2).

The K1 sample is a coarse- to medium-grained equigranular quartz diorite - tonalite from the northern edge of Aishihik Batholith. Its $\delta^{18}O$ values are +5.9 and +8.8°/₀₀ for whole-rock and quartz respectively. Petrographic observation of ferromagnesian phases indicates a high probability of interaction with altering fluids.

The porphyritic biotite quartz monzonite suite is represented by two samples; P1 and P2. Their whole-rock $\delta^{18}O$ are respectively +8.0 and +7.7°/₀₀, which is normal for

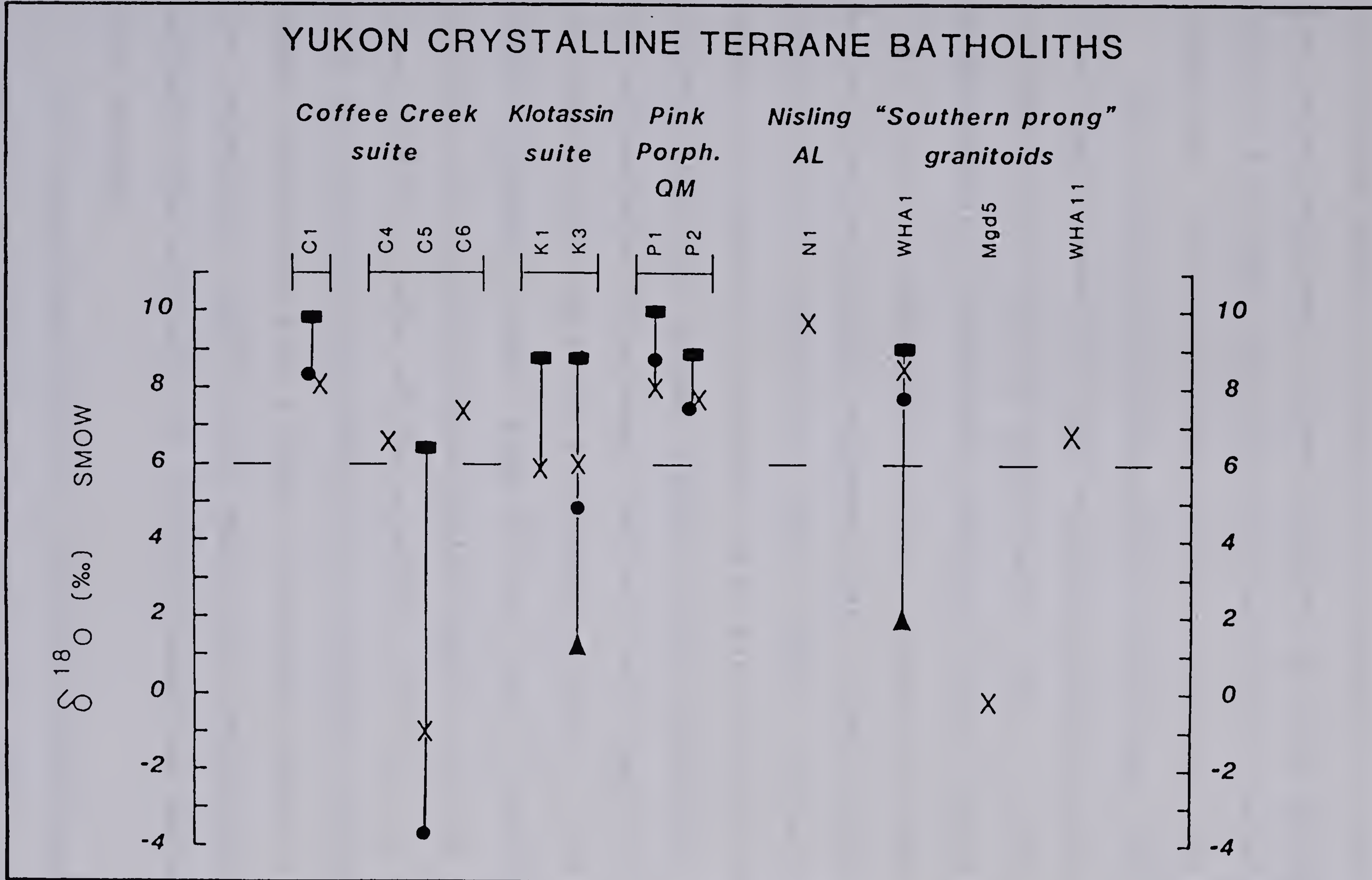


Figure 2-2 <u>$\delta^{18}O$ results for the Yukon Crystalline Terrane granitoid rocks.</u> The sample numbers are given at the top of the diagram. The dashed line is the lower limit of normal primary $\delta^{18}O$ values for granites. The symbols represent: rectangle → quartz, dot → feldspar, X → whole-rock, and triangle → biotite.

Cordilleran granitoids. Even though ferromagnesian (hornblende mostly) alteration is present to a great extent, the relatively small volume occupied by these phases in the rock (<10 %) does not allow for lowering of the primary whole-rock $^{18}O/^{16}O$ ratio by more than about 0.1 to 0.2°/₀₀.

Two volumetrically extensive rock suites, the Coffee Creek quartz monzonite and the Nisling Range alaskite, are represented by only one sample each (resp. C1 & N1). The C1 sample has an $^{18}O/^{16}O$ ratio similar to the P1 and P2 samples (+8.0°/₀₀). Isotopic partitioning in quartz-feldspar pairs as well as petrography suggests that it has a primary oxygen-isotope composition. The N1 sample, on the other hand, is a miarolitic alaskite that was collected from an outcrop face that showed intense surface weathering. This is not an isolated occurrence, but is a very characteristic feature of this rock suite (Goodfellow *et al.*, 1978) in the region. Because of the large (0.5 to 1 cm +) interconnected cavities, the weathering permeates deeply into the outcrop face. The $\delta^{18}O$ value of the alaskite (+9.7°/₀₀) is rather high and could be explained by two different processes; magma interaction with a high ^{18}O crustal reservoir and/or low temperature (≤25°C) clay weathering. Petrography suggests that at least the second process was operative in this sample.

In the southern prong of the Yukon Crystalline Terrane, three granite samples have been collected. WHA 1 is a coarse-grained leucocratic biotite granite of probable

Tertiary age. It is part of a series of small, 50-100 km^2 subcircular to elongated intrusions outcroping in the Yukon Crystalline Terrane and the Coast Plutonic Complex. The $\delta^{18}O$ of quartz (+9.0°/oo) and feldspar (+7.7°/oo) are normal values for this type of rock. The measured whole-rock $\delta^{18}O$ of +8.5°/oo indicates that this sample has not undergone any significant sub-solidus ^{18}O exchange with low $\delta^{18}O$ fluids.

The second sample, WHA 11, is a granodiorite boulder that was collected from an outcrop of the Jurassic Laberge conglomerate near Whitehorse. K-Ar dating suggests an age of about 140 Ma significantly older than the norm in the Whitehorse area (Morrison *et al.*, 1979). It is highly altered and its $\delta^{18}O$ of +6.7°/oo is very similar to the Klotassin results. Old boulders of this sort occur throughout the northern part of the Yukon Crystalline Terrane and are interpreted as being eroded parts of the batholiths of the area. As such, the sample was included in the data set of this region and not in the Whitehorse results.

The last granitic sample to be reported on from this region (Mgd-5) was collected on the south shore of the west end of Bennett lake. It is a gneissic, feldspar porphyroblastic granodiorite of possible Triassic age (Morrison *et al.*, 1979). Outcrop samples from this locality are generally of poor quality because of pervasive shearing and alteration that gives the feldspar megacrysts a chalky look. Epidote and chlorite veinlets are locally abundant.

Petrographical observations confirm the presence of secondary minerals, which comprise a major fraction (approx. 35 to 40%) of the rock. The $\delta^{18}O$ of $-0.2°/_{oo}$, obtained for Mgd-5, is in good agreement with the altered nature of the rock, noted above.

C. Coast Plutonic Complex granitoid rocks

Some 24 samples from this region were analysed, of which all but seven are from the Mesozoic coast plutons (Table 2-3) . Rock powders and in some instances, rock samples were made available by previous workers (Morrison *et al.*, 1979, Bultman, 1979). Limited additional sampling was done to cover the area more uniformly (figure 2-3). The whole-rock average is $+6.0°/_{oo}$, a result somewhat lowered by alteration processes. Statistically, altered rocks were fewer in number in this collection than in the ones from preceding sections. It is doubtful that this is a simply the result of biases in sampling technique since hydrothermally altered rocks often show no macroscopic and even microscopic clue to their nature. Excluding hydrothermally affected rocks from the mean calculation increases its value to $+7.5°/_{oo}$, typical of primary $\delta^{18}O$ in calc-alkaline batholiths.

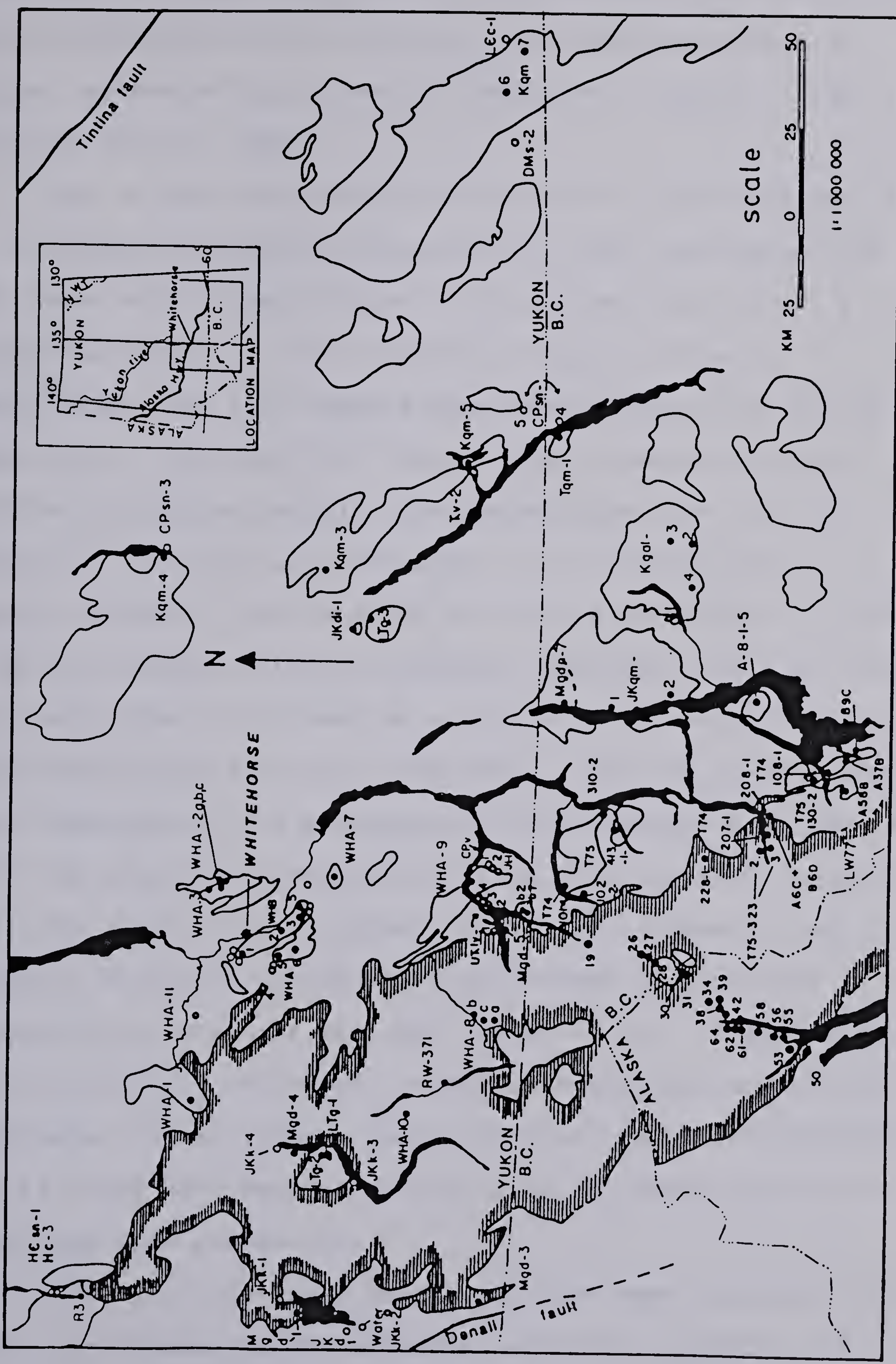
No systematic variation in whole-rock $\delta^{18}O$ results is seen in the coast mountains as opposed to the fairly regular increase in ^{18}O, eastward across the Peninsular Range batholith in southern California (Taylor & Silver, 1978). As

Table 2-3 : $\delta^{18}O$ of the Coast Plutonic Complex batholiths

Sample	Rock Type	Whole Rock	Qtz	Fp	Bio	Hbl
Llewellyn Inlet porphyritic granodiorite						
T74 108-1	GD	7.3	8.5	7.2		
T75 130-2	GD	-0.6	8.5	-0.8		
LW77 B9C	GD	7.0				
LW77 A37B	GD	9.3				
Mount-Caplice granodiorite						
LW77 A58B	GD	7.5				
Florence Range and Mount-Lawson granitoids						
T74 208-1	GD	7.8				
LW77 A6C	GD	7.3				
LW77 B6D	GD	7.5	8.7	7.5 ph	(7.8 Grm)†	
T74 207-1	GD	-0.8	7.1	-4.0		
T75 323-2	Di	5.7				
T75 323-3	Gr	8.4	9.3	7.4		
T74 228-1	GD	7.2	9.5	7.1		
Whitehorse area Coast Plutonic rocks						
WHA 8a	GD	5.9	8.6	6.5	1.9	
WHA 8b	QM	6.6				
WHA 8c	Di	3.8	8.6	0.4		4.5
RW 371	QD	2.5				
WHA 10	GD	5.7	9.0	6.0	1.9	6.1
R2	GD	3.6	7.7	2.8		
R3	GD	9.4				
lTg-1	Gr	-0.6				
lTg-2	Gr	9.2	9.5	8.5		
Mgd-4	GD	9.2				
Mgd-3	GD	7.1	11.3	8.7		
Mgd-1	GD	8.3				

† $\delta^{18}O$ value for the groundmass feldspars, ph = $\delta^{18}O$ of the phenocrysts.
All values are in permil (SMOW). Di represents diorite and Gr - granite, all other abbreviations as before.

Figure 2-3 <u>Sample locations in the southern Yukon and northern British Columbia.</u> Dots represent plutonic samples, open circles are country rocks. Plutons are outlined in ink. The Coast Range Batholith is partly shaded. Dark areas are large lakes or bays. Samples # 64 through 26 and B2, B7 in southeast Alaska (lower left-hand corner of diagram) are from a previous study by Magaritz and Taylor (1976a). All others are from this study. Not all sample locations in the Whitehorse Trough are shown in this figure.


Tintina fault
WHITEHORSE
YUKON
B.C.
ALASKA
Denali fault
N
LOCATION MAP
scale
KM 25 0 25 50
1:1 000 000

expected, a strong similarity exists between the results of this study and similar studies of the Coast Mountains in other regions of North America (Magaritz & Taylor, 1976a, b, Criss & Taylor, 1982).

One of the oldest dated intrusions in the Atlin area is a K-feldspar porphyritic granodiorite that outcrops on the northern and southern shores of Llewellyn Inlet, B.C.. A large variation in $^{18}O/^{16}O$ ratios (figure 2-4) was discovered from four samples that are assumed to be of the same pluton (Bultman,1979) though their textural features differ quite considerably. The rocks range from -0.6 to +9.3‰. The most altered sample, as evidenced by a quartz-feldspar fractionation of +9.3 is T75-130-2. It comes from the southern side of Cathedral Mountain, which on the northern side is intruded by a younger (Cretaceous?) hornblende quartz diorite (T75-208-1). The two samples above are separated by 5-6 kilometers. The hornblende K-Ar age in T75-130-2 is reset from about 215 Ma (date found in sample T74-108-1) to 175 Ma (Bultman, 1979). It is thought that the younger intrusion to the north has caused this partial resetting of the K-Ar age. The timing of the ^{18}O depletion of T75-130-2 is not known; it may have occurred during the intrusion of the younger quartz diorite (142 Ma) (T75-208-1) or it could have happened during later plutonic activity in the area (see chapter 3).

LW77 B9C is another sample from the same intrusion. It intrudes Permian or Pennsylvanian andesites, basalts and

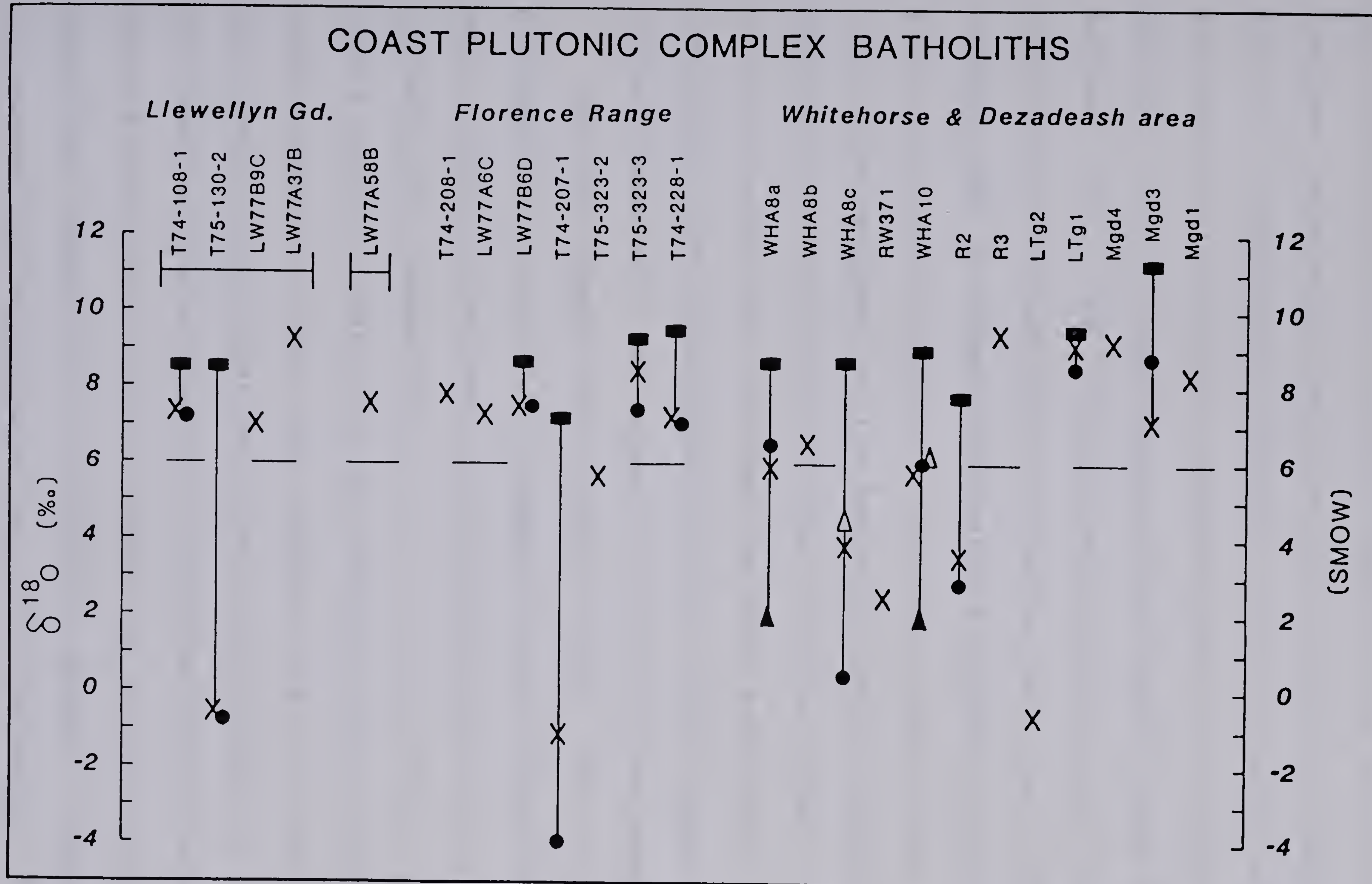


Figure 2-4 $\delta^{18}O$ results for the Coast Plutonic Complex granitoid rocks. The symbols are: rectangle → quartz, dot → feldspar, X → whole-rock, filled triangle → biotite, and open triangle → hornblende. The dashed line represents a benchmark isotopic value for primary $\delta^{18}O$ of granites.

pyroclastic rocks. The sample is moderately propylitised and its $\delta^{18}O$ (+7.0°/₀₀), although slightly low, does not indicate large scale depletion. This holds true especially when it is compared to T75-108-1 (+7.3°/₀₀) which exhibits equilibrium fractionation between quartz and feldspar ($\Delta^{18}O$ qtz-fp = +1.3°/₀₀) and has been described as fresh by Bultman (1979). The ^{18}O differences between these two samples could be ascribed to relative positions in the magma chamber; B9C is close to the eastern granodiorite/volcanic rock contact while 108-1 is located at least 500 meters within the periphery of the batholith and is topographically (perhaps also structurally) lower than the former. Such a position, would have isolated the 108-1 sample more thoroughly from hydrothermal activity, often present at contact zones and other areas of structural weakness.

The last sample from Llewellyn Inlet, LW77 A37B, is enriched in ^{18}O (+9.3°/₀₀) compared to the other three rocks above and also from what is considered normal for a rock of granodioritic composition, *i.e.* +7.0 to +8.5°/₀₀. It comes from a central location in the batholith and is located higher relative to the other samples from this body. Fractional crystallisation is not a valid cause of this enrichment since no change in rock mineralogy is noted and because the size of the ^{18}O spread (up to 2.3 °/₀₀) is beyond what separation of crystals at magmatic temperatures can accomplish (Matsuhisa *et al.*, 1973, Matsuhisa, 1979, Muehlenbachs & Byerly, 1982). Isotopic exchange with the

medium to high grade metamorphic country rocks in the immediate area (mafic schists and gneisses) could account for this enrichment.

The Mt-Caplice granodiorite sample (LW77 A58B), situated just a few kilometers west of the Llewellyn intrusive, has a slightly older age than the latter (230 vs 215 m.y) It is slightly sericitised and shows a normal +7.5 whole-rock $\delta^{18}O$. Significant K-Ar resetting of hornblende and biotite to 100 and 50 Ma respectively (R. Armstrong, unpublished data) has occurred in this sample. However, the resetting event does not seem to have affected the whole-rock $\delta^{18}O$ of the intrusive.

Further north, a series of seven samples from the southern extension of the Taku arm of Tagish Lake were analysed. They form a compositionally and texturally heterogeneous group ranging from granodiorite and tonalite to quartz monzonite. Three of these, near the mouth of the Wan river (T75-208-1, LW77 A6C & B6D) show little variation in whole-rock $\delta^{18}O$ (+7.8, +7.3, +7.5‰). Petrographic observation of A6C, a medium grained equigranular hornblende-biotite granodiorite, shows very fresh feldspar and biotite grains but very well crystallised epidote (sensu stricto) and clino-zoisite clusters after hornblende and also some chlorite. Volumetrically the secondary phases are minor when compared to the fresh phases of the rock. This sample is altered to a degree and its ^{18}O concentration is the lowest of the three samples above. When it is compared

with the adjacent B6D sample which exhibits an equilibrium $\Delta^{18}O$ (qtz-kfp) value of 1.2‰ and a very similar whole-rock $\delta^{18}O$ of 7.5‰, the isotopic data suggests that A6C has not been depleted significantly by the alteration.

Three other samples, a few kilometers west of the above, show very distinct compositions. A coarse grained biotite granite (T75-323-3) is the youngest rock (59 Ma), according to intrusive relationships and K-Ar dating (Bultman, 1979). Its $\delta^{18}O$ (+8.4‰) is also the highest of the seven samples in the area. The two surrounding rocks (within 1-2 km of each other) are depleted in ^{18}O, T74-207-1 to a large degree (-0.8‰) and T75-323-2 less so (+5.7‰). Thermal resetting of the K-Ar clock in the former sample to 64 Ma (Bultman, 1979), probably dates the hydrothermal event that affected both altered samples and points to the youngest granite (T75-323-3) as the cause of this effect.

The last of the seven samples in the Taku Arm area, is a hornblende rich tonalite (T75-228-1) from Mt-Lawson. It intrudes coast metamorphic rocks on its eastern flank. Its $\Delta^{18}O$ quartz-feldspar value of 2.4 is within the 1 to 2.5‰ magmatic spread for these two minerals. The presence of 5% epidote in thin section does not seem to have considerably affected the whole-rock $\delta^{18}O$ value (7.2‰) for this sample.

In the Whitehorse map area, five rocks from the Coast Plutonic Range have been studied. Their composition is on

the more mafic end of the granite spectrum, ranging from pyroxene diorite to quartz monzonite.

Three samples from the Mt-Ward intrusion (WHA 8a, 8b, 8c) have relatively low $\delta^{18}O$ values (+6.6, +5.9, +3.8°/oo). The most ^{18}O depleted rock (8c) is an augite rich, hornblende porphyritic diorite. Hornblende phenocrysts are poikilitic with inclusions of clinopyroxene, secondary chlorite, feldspar and quartz. Augite crystals are well formed and fresh with only traces of chlorite present on the grain surface. Feldspar is also generally clear of secondary minerals except in localised clusters where sericitisation is moderate. The $\delta^{18}O$ mineral results do not totally agree with the above observations as they show low-^{18}O feldspar (+0.4°/oo) which is isotopically reversed when compared to hornblende in the rock ($\Delta^{18}O$ pfp-hbl = -4.1°/oo). Such a large isotopic depletion in depletion in plagioclase without petrographic signs of secondary alteration has been observed previously in the Skaergaard intrusion (Taylor & Forester, 1979) where low ^{18}O fluids must have interacted at much higher temperatures (500-600°C) than is common in granitic systems (200-400°C).

The WHA 8a sample, a coarse grained granodiorite, has been significantly less affected by alteration than the previous sample. Biotite is the only mineral that shows petrographic signs of chemical disequilibrium as progressive vermiculitisation from the edges inward is observed throughout the thin section. Quartz, feldspar and biotite

give $\delta^{18}O$ values of +8.6, +6.5 and +1.9°/oo respectively, which indicates isotopic disequilibrium of the biotite.

Further west, a sample of equigranular pyroxene-bearing biotite quartz diorite (WHA 10) was collected from the mountain overlooking Rose Lake. Its whole-rock $\delta^{18}O$ of +5.7°/oo is similar to the Mt-Ward granite results. Feldspar and pyroxene are isotopically reversed; a clear sign of disequilibrium. Petrographically, plagioclase (An_{55}) is free of alteration while augite is partially transformed to uralite , chlorite and to a lesser degree serpentine minerals. The $\delta^{18}O$ of these altered pyroxene clusters (+5.2°/oo) are sufficiently close to the normal magmatic pyroxene range (+5.8 to +6.5°/oo) to suggest very little interaction with low ^{18}O fluids. Feldspar on the other hand, is 3 °/oo lower than quartz, which is well below the equilibrium partitioning for these minerals at magmatic temperatures (Matsuhisa *et al.*., 1979). Obviously feldspar has isotopically retrograded at lower temperatures, after crystallisation of the rock.

In the Dezadeash area, samples Mgd-1 and Mgd-3 were taken from the edge of the Coast Plutonic Batholith. Mgd-3 is a coarse-grained biotite hornblende quartz diorite ($\delta^{18}O$ = +7.1°/oo). The $\Delta^{18}O$ (qtz-fp) equals +2.5°/oo, just within the the magmatic ^{18}O range for these minerals and the $\delta^{18}O$ of quartz is a very high +11.3°/oo. Slight to moderate sericitisation is seen along cleavages and fractures in the cores of the zoned plagioclase crystals. Biotite is

moderately vermiculitised and to a lesser degree chloritised. Hornblende is generally fresh with only traces of epidote present after amphibole.

Mgd-1 is also quartz dioritic in composition ($\delta^{18}O$= +8.3°/₀₀) with very similar modal proportions of minerals as Mgd-3. Clinopyroxene crystals are rimmed with uralitic amphibole as in sample WHA 10. All other phases in the rock are fresh.

Three samples for the western shore of Kusawa Lake show strikingly different isotopic features. Samples lTg-1; a coarse-grained porphyritic biotite quartz monzonite and Mgd-4, a biotite granite, are mineralogically similar but differ in ^{18}O (-0.6 and +9.2°/₀₀, respectively). Perthitic feldspars and microcline are moderately kaolinitised in both rocks while chlorite replaces biotite to a greater extent (>50%) in Mgd-4 than in lTg-1 (<15%); in opposition to what would be expected based on ^{18}O data. A younger, possibly Early Tertiary, coarse-grained hornblende-biotite granite (lTg-2) intrudes the granite-quartz monzonite suite above and has a high +9.2°/₀₀ $\delta^{18}O$ value; normal for this type of rock. A 10°/₀₀ gradient between samples lTg-1 and lTg-2 along a distance of less than 10 metres is unusual. The contact between these two rocks occurs in a structurally weak zone where recent (Pleistocene or Holocene) sediments were deposited and through which a creek is flowing. It is therefore not known if this gradient is gradual over the distance covered or if the $^{18}O/^{16}O$ ratios change

dramatically over very short distances (<1m) near the contact. A similar large ^{18}O gradient has been observed in the faulted and sheared contact zones of the Whitehorse batholith between depleted skarns (-5 to -9°/₀₀) and the pluton (+8 to +6°/₀₀) by Morrison (1981). Very high water fluxes must have occurred within narrow bands of rock near the Whitehorse intrusion. This could explain the depletion of sample lTg-1 and its proximity to the younger intrusion (lTg-2) accounts for the lack of secondary minerals.

Finally, two samples were taken from Ruby Range granodiorite. Both are biotite-bearing, medium-grained granodiorites, sample R2 being hornblende-rich, while R3 having no amphibole. Clay after feldspar is present in both samples but is slightly more abundant in sample R2 ($\delta^{18}O$ =+3.6°/₀₀). The clay minerals in R3 are situated in the spaces between mineral grains rather than throughout the minerals themselves. This produces a rock that in hand specimen is very brittle. Chloritisation of minor biotite is more extensive in R3 (≤50%) than in R2. Sample R3 ($\delta^{18}O$ = +9.4°/₀₀) was collected at Otter Falls on the Aishihik Lake Road where all exposures of this rock type consisted of very friable, rust stained material, probably the result of low temperature weathering. Both samples are only slightly altered. The low $\delta^{18}O$ result for R2 makes it similar to other isotopically exchanged but not mineralogically altered samples (*i.e.* WHA 10, 8c, and 2). In the case of R3, the minor amounts of clay minerals present in the rock may have

increased its whole-rock $\delta^{18}O$ by not much more than $0.5^{\circ}/_{oo}$.

D. Atlin Terrane batholiths

Atlin granitoid rocks intrude Upper Paleozoic oceanic sedimentary rocks (chert, carbonate and argillite) that have been reversely thrusted over the younger rocks of the Whitehorse Trough. Their size is an order of magnitude larger than their counterparts in the Whitehorse area (>500km^2 vs. 100 km^2) and their isotopic and chemical compositions are more enriched in ^{18}O and silica, respectively. The above observations make the Atlin rocks more akin to the Omineca granites, to the east, than to those discussed thus far.

The unweighted average whole-rock $\delta^{18}O$ for the six batholiths sampled † is $+8.3^{\circ}/_{oo}$ (Table 2-4), at least $2^{\circ}/_{oo}$ richer in ^{18}O than the averages calculated for the terranes to the west. Only one of the 9 samples analysed, Tqm-1, is significanty depleted in ^{18}O ($\delta^{18}O = -2.5^{\circ}/_{oo}$). Excluding this value from the mean calculation yields an average of $+9.7^{\circ}/_{oo}$; a result much more representative of the Atlin granites isotopic signatures based on rock volume. It must be pointed out that even though only one sample is isotopically depleted, this does not necessarily mean that less interaction has occurred between groundwater and rock in this area than in the western terranes. An ^{18}O enriched fluid cycling through a hydrothermal system could also

---

† See figure 2-3 for the approximate locations.

Table 2-4 : $\delta^{18}O$ of the Atlin Terrane batholiths

Sample	Rock Type	Whole Rock	Qtz	Kfp	Bio
Suprise Lake batholith					
Kgal-1	Al	12.2	12.4	10.3	
Kgal-2	Al	10.0	11.2	9.3	6.1
Kgal-3	Al	9.2	9.5	9.0	5.3
Kgal-4	Al	11.3	12.6	10.3	5.6
Other granitic bodies					
JKqm-1	QM	11.4			
JKqm-2	QM	9.3			
Mgdn-7	Gdn	9.6			
A 8-1-5	GD	8.4			
lTg-3	Gr	7.7			
JKdi-1	Di	6.8			
Tqm-1	QM	-2.6			

All values are in permil (SMOW). Gdn = granodiorite gneiss, Di = diorite, other abbreviations as before.

produce this feature.

The most interesting suite sampled in the area is comprised of four samples from the Surprise Lake Batholith (Kgal-1, -2, -3, -4). The rocks are part of an ilmenite series granite-alaskite with very peculiar chemical characteristics. It is mineralised in Mo-Zn-W-Sn and is enriched in U, Th, F, Sn, Rb, Li, Y, Nb, Zn, Pb and REEs relative to average Ca-poor granites (B. Ballantyne, personal communication).

The first sample (Kgal-1) was taken from an uncomformable contact with recent olivine basalts at the Ruby Creek exposure. Structurally, the sample is close to the now eroded paleocontact between the granite and the sediments of the Cache Creek group comprised of chert, greenstone and carbonate. Large exposures of ultramafic rocks are also seen throughout the area. K-feldspar phenocrysts have a milky-white color and show a moderate degree of kaolinisation in thin sections of sample 1. Plagioclase and biotite grains in the matrix are also altered and a limonitic staining is apparent on the outcrop. Its whole-rock $\delta^{18}O$ of +12.2°/₀₀ is the highest recorded in the area. The quartz-alkali feldspar fractionation of +2.0 permil (figure 2-5) is only slightly greater than the normal $\Delta^{18}O$ of +1.5°/₀₀ for granites (O'Neil & Taylor, 1967, Blattner & Bird, 1974, Matsuhisa *et al.*,1979) Such a small departure from equilibrium values is somewhat surprising considering the altered state of the constituent minerals.

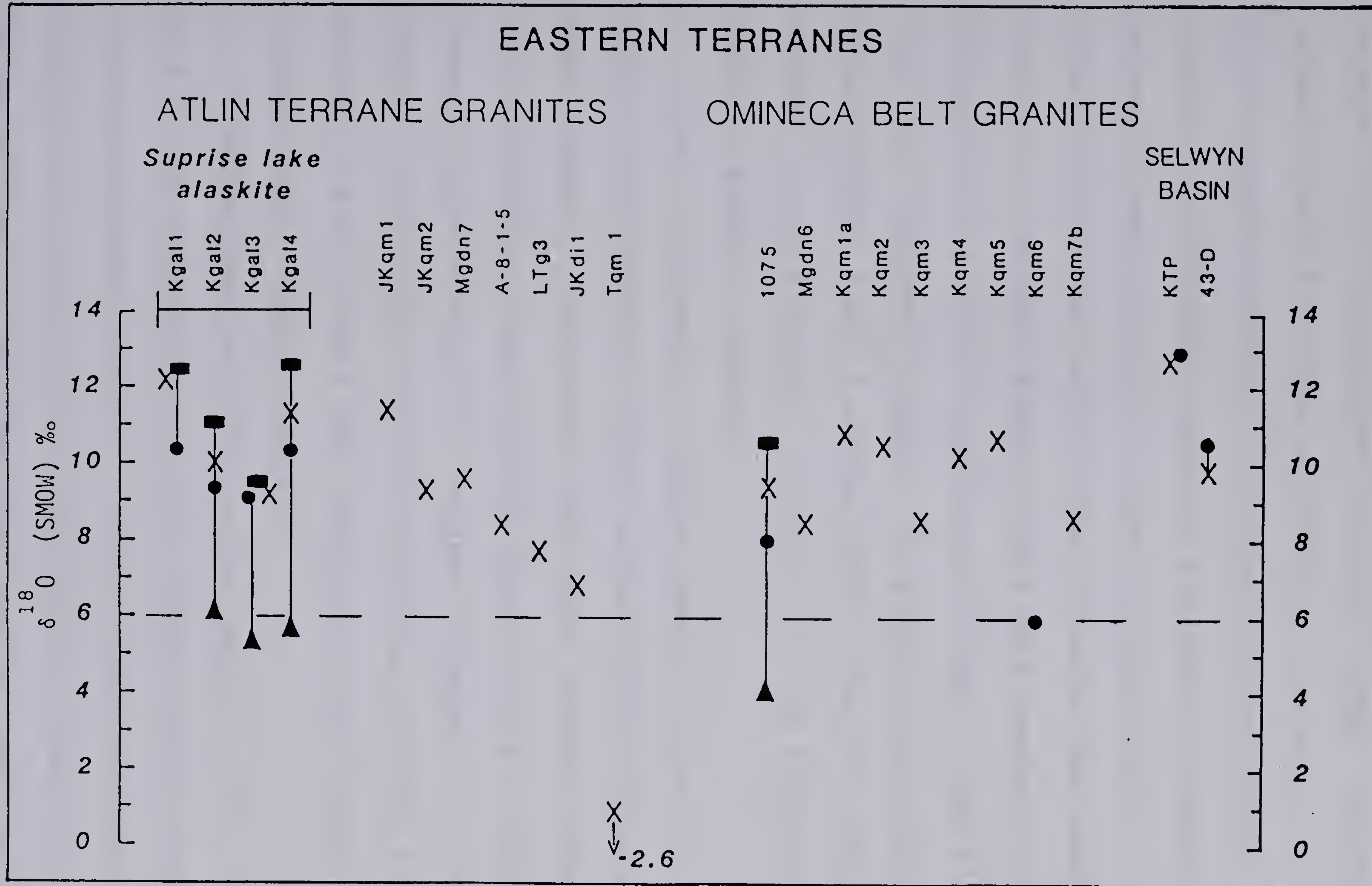


Figure 2-5 <u>$\delta^{18}O$ results for the eastern terranes.</u> Symbols: rectangle → quartz, dot → feldspar, X → whole-rock, and triangle → biotite. The dashed line represents the isotopic value of the mantle and serves as a benchmark for primary ^{18}O compositions.

Sample Kgal-2 is a geologically problematic phase because of its hornblende-rich nature. Trace element data suggests that it is less evolved than the main biotite-bearing phase of the Surprise Lake batholith or possibly, a different intrusion altogether (Ballantyne, personal communication). It was collected from a structurally deeper part of the intrusion than samples Kgal-3 or 4. Mineral phases in the rock, except quartz, are altered in an argillic fashion as in Kgal-1. The $\delta^{18}O$ of +10.0°/₀₀ obtained for Kgal-1 is significantly depleted versus samples 1 and 4 of the pluton. The $\Delta^{18}O$ (qtz-Kfp) though, is quite similar (+1.9°/₀₀) to the fractionation found in sample 1 (above).

Kgal-4 represents a typical sample of the unmineralised phase of the batholith from a structurally high zone in the magma chamber, possibly the roof zone. Mineral phases in this specimen are less altered than in the two rocks above, though clay minerals still replace feldspars to a small degree (<15%). Both the $\delta^{18}O$ (+11.3°/₀₀) and the $\Delta^{18}O$ qtz-Kfp (+2.3°/₀₀) are very similar to the isotopic compositions of Kgal-1 and 2.

The last sample of this suite, Kgal-3 ($\delta^{18}O$ = +9.2), is part of the molybdenum-hosting, greisen veined, stockwork of the Surprise Lake alaskite. It was collected from a mineralised fracture face. Clay mineral formation is more pervasive than in the other Kgal samples. Sericite on plagioclase and K-feldspar crystals is also present in the

sample. The hand specimen is highly rust-stained throughout, intensifying in color towards the fracture face. The $\Delta^{18}O$ between quartz and K-feldspar (+0.5°/oo) is too small to be magmatic (O'Neil & Taylor, 1967) such that an explanation for the relative enrichment of feldspars and biotite has to be found. A depletion of ^{18}O in quartz is an unlikely explanation for the small mineral fractionations observed, since isotopic reequilibration has been shown to occur much more readily in feldspars at all post-solidus temperatures especially in the presence of salt-rich fluids (O'Neil & Taylor, 1967, Cole & Ohmoto, 1976, 1979, Criss & Taylor, 1982).

Two samples from the Mt-McIntosh batholith were collected on the eastern shore of Atlin Lake. Jkqm-1 is a sample of melanocratic medium- to coarse-grained hornblende-biotite quartz monzodiorite with a $\delta^{18}O$ of +11.4°/oo. Moderate sericitisation and chloritisation are present. Uralitic amphibole almost completely replaces primary clinopyroxenes and is itself replaced to a degree by epidote and chlorite.

The second sample, JKqm-2, is a medium grained monzonite with more pronounced alteration ($\delta^{18}O$ =+9.3°/oo). Albite twinning is destroyed in over 70% of the plagioclase grains by the growth of sericite. All other phases except quartz and uralite have been affected by the alteration process. If one takes the $\delta^{18}O$ of the previous sample as representing close to a primary isotopic composition for the

batholith, then JKqm-2 must have been depleted in ^{18}O by at least $2^{\circ}/_{oo}$. It is significant to note that rocks of the other western terranes that show a mineralogically similar degree of hydrothermal alteration, have $\delta^{18}O$ values ranging from $-1.0^{\circ}/_{oo}$ to about $+2^{\circ}/_{oo}$.

Mgdn-7, an epidioritic pyroxene-hornblende rich intrusion on Mt-Hitchcock has the same petrographic features described above. Its $\delta^{18}O$ of $+9.6^{\circ}/_{oo}$ is also very close to the ratio from the altered sample of the Mt-McIntosh batholith. Although sometimes mapped as separate intrusions, it is probable that the three preceding samples are part of the same intrusive suite.

The last four samples from the area, all crop out along its structural margin. On the western side of the region, a tonalitic to granodioritic, biotite-hornblende-bearing, circular intrusion is found on Theresa Island ($\delta^{18}O$ = $8.4^{\circ}/_{oo}$). It intrudes along the trace of the Nahlin fault into Paleozoic ultramafic and mafic volcanic rocks. It is very fine-grained which suggests that it was emplaced epizonally.

On the north-eastern side of the Atlin Terrane, another subcircular hornblende-biotite granite ($\delta^{18}O$ =$+7.7^{\circ}/_{oo}$) intrudes Permian volcanic rocks and Late Paleozoic Cache Creek sedimentary rocks. lTg-3 is not affected much by alteration with only the most calcic cores of zoned plagioclase grains being slightly sericitised. An older serpentinised uralitic epidioritic stock just 0.5 km north

of the granite has a relatively normal $\delta^{18}O$ of $+6.8°/_{oo}$ for this type of rock. This result is in disagreement with petrographic observations indicating strong secondary recrystallisation.

The last sample reported on from the Atlin region is Tqm-1 which is situated on the western shore of Teslin Lake. As with the two previous intrusions, this quartz diorite is a small (<50 km^2) circular pluton emplaced in a zone of structural weakness; the Teslin Lineament. Its isotopic ratio is the lowest encountered in the area and in this entire study ($-2.6°/_{oo}$). The intrusion was probably emplaced in Tertiary time, after faulting along Teslin Lake had occurred. The faulting must have provided very efficient pathways for groundwaters to infiltrate and deplete this pluton to such an extent.

E. Omineca Crystalline Belt granites

This geological province, although important in the context of the Cordilleran Orogeny was not the focus of this study. Only ten samples were obtained from previous workers or were collected in the field and analysed. The purpose of the analyses was to obtain a regional picture of the oxygen isotope ratios. The compositionally restricted siliceous nature of the granites and the presence of muscovite in some rocks suggests an anatectic origin for these batholiths. Similar degrees of alteration have affected the Omineca rocks and the granitoid rocks from the other sections. Table

2-5 shows the whole-rock and mineral $\delta^{18}O$ results for this region.

The two samples from the Selwyn Basin (figure 2-3) are hypabyssal two-mica quartz monzonites containing feldspar phenocrysts. KTP ($\delta^{18}O$ = +12.7°/oo) intrudes meta-argillaceous carbonate-bearing strata and 43-D ($\delta^{18}O$ = +9.8°/oo) is emplaced in mostly volcanic rocks near the Tintina Fault.

The remaining eight samples are part of the group from the western side of the Tintina Trench. The average $\delta^{18}O$ is +9.6°/oo, very close to the $\delta^{18}O$ mean value from the Atlin terrane. There is a bimodal distribution in the data set of table 2-5 with one group ranging from 8 to 9°/oo and the other; 10-10.8°/oo. This is possibly due to the geological locations of the samples; the higher $^{18}O/^{16}O$ ratios were obtained from rocks near contacts with isotopically enriched carbonaceous, argillaceous and quartz-rich country rocks. The lower ^{18}O group are rocks taken from the center of large intrusions or near contacts with more permeable and more ^{18}O poor volcanic country rocks.

Summary

In summary, examination of the data presented in this chapter yields the folowing observations;

1. Low-^{18}O granitic rocks characterised by isotopic mineral disequilibrium are common in the three western terranes of this study (CPC, WT, YCT). In the Whitehorse Trough

Table 2-5 : $\delta^{18}O$ of Omineca Belt and Selwyn Basin granites

Sample	Rock Type	Whole Rock	Qtz	Fp	Bio
Selwyn Basin granites					
KTP	QM	12.7		12.8	
43-D	QM	9.8		10.6	
Ominica Crystalline Belt granites					
1075	Gr	9.4	10.6	7.9	4.0
Mgdn-6	Gdn	8.4			
Kqm-1	QM	10.8			
Kqm-2	QM	10.5			
Kqm-3	QM	8.0			
Kqm-4	QM	10.2			
Kqm-5	QM	10.7			
Kqm-6	QM			5.9	
Kqm-7b	QM	8.6			

All results are in permil (SMOW).

area these rocks predominate in number and volume over isotopically normal granites.

2. Identical alteration patterns do not necessarily produce similar ^{18}O signatures; in the eastern terranes, altered rocks tend to preserve their original high $^{18}O/^{16}O$ ratios, while further west, altered rocks generally are depleted in ^{18}O and have low $^{18}O/^{16}O$ ratios.
3. A good correlation exists between the degree of mineral alteration and the amount of isotopic disturbance found in rocks of the three western terranes (WT, CPC, YCT). Exceptions to this are found mainly in mafic rock varieties, which may have been altered at temperatures beyond the stability range of hydrous minerals.

It is clear from the oxygen isotope data above that hydrothermal conditions must have been broadly similar in the western parts of the study area to produce the same alteration patterns in the three western geological provinces. This section has concentrated primarily on describing the geological setting of the plutons, their mineralogical features, their general isotopic features and has tried to suggest, briefly, some of the causes of the observed isotopic patterns. Since a great many of the samples have been depleted in ^{18}O by isotopic exchange with a light oxygen source, the next chapter will concentrate on the conditions in which this occurred. The Whitehorse area granitoid rocks are most suited for this investigation since they are more densely-sampled than rocks from other regions

and show some of the more depleted examples. As such, they will be the focus of the discussion in the next chapter.

III. HYDROTHERMAL ALTERATION

A. Review of general features

Granitic rocks throughout the world have been well studied by numerous workers. The primary ^{18}O range for these rocks is about +6.0 to +14.0 ‰. It includes rocks that have generally obtained their isotopic signatures through primary igneous processes, regardless of the source of the magmas. The lower limit of the range is the inferred oxygen isotopic composition of the earth's mantle (≅ +6.0‰) and its primary derivatives such as MORBs. Differentiated plutonic rocks tend to be ^{18}O enriched by at least 1.5 to 2‰ relative to the mantle and mafic volcanic rocks. As evidenced by lunar rocks which cluster around the mantle $\delta^{18}O$ value regardless of their chemical composition and mode of emplacement, the ^{18}O enrichment in plutonic rocks on earth is probably the result of sedimentary cycling and metamorphism through time (Epstein & Taylor, 1971, Longstaffe & Schwarcz, 1977, Longstaffe, 1979). Typical $\delta^{18}O$ values for large granitic batholiths along plate margins is +7.0 to +9.0 ‰ , while those from more continental locations are usually richer in ^{18}O and may range up to +14‰. Their increased concentration of the heavy isotope often reflects their derivation from or interaction with high ^{18}O sedimentary and metamorphic rocks. Igneous processes have not been found to produce significant depletions in the $\delta^{18}O$ of magmas. Theoretically, fractional

crystallisation and removal of ^{18}O rich phases from a melt would result in a lowering of the whole-rock $\delta^{18}O$ value of late differentiates, but normal crystal fractionation tends to increase the $\delta^{18}O$ of the melt as ^{18}O poor minerals such as olivine and pyroxene are segregated from the liquid. In any case, large changes in ^{18}O composition cannot be accomplished in such conditions (>800°C) since fractionation factors between minerals and melts approach unity at magmatic temperatures (Matsuhisa *et al.*, 1973, Matsuhisa, 1979, Muehlenbachs & Byerly, 1982). To explain large scale lowering of whole-rock $\delta^{18}O$ values often seen in epizonal plutons, lower temperature conditions have to be inferred.

Water, as a large reservoir of oxygen in the crust, is usually involved in large scale ^{18}O exchange between rocks, and can easily deplete their isotopic composition, given appropriate geological conditions. As was shown by Craig (1961b, 1963), Dansgaard (1964) and Friedman and others (1964), the ^{18}O and the deuterium (D) composition of surface and groundwaters becomes progressively depleted with respect to ocean water at higher latitudes and/or altitudes. At one extreme, surface precipitation on Antartica has a $\delta^{18}O$ of -56 (δD = -430 ‰), while rain in oceanic regions is essentially identical in isotopic composition to sea water. In the elevated mountainous regions of the North American Cordillera, meteoric waters range from about -10 in the south to to -24 ‰ in the north.

In this study, one sample of well water was taken from the Dezdeash map area, near Beloud Post, Yukon. Its $\delta^{18}O$ value of -21.4°/₀₀ is consistent with its geographical position (60 ° North) and its altitude (over 720 m above sea level). Given its depleted nature, meteoric groundwater in the area is the most likely low-^{18}O reservoir available to produce the depleted $\delta^{18}O$ values of the granitic rocks described in the previous section.

Meteoric-hydrothermal water-rock interactions have been well documented in the Cordilleran region (Sheppard *et al.*, 1969, 1971, Taylor, 1971, 1974a, b, 1977, Forester & Taylor, 1972, 1974, 1977, Magaritz & Taylor, 1976a, 1976b, Criss & Taylor, 1982). These intrusions initiated hydrothermal systems involving large amounts of water in a convective circulation regime. Fluids mostly move through the rock pores and fractures that develop near zones of weakness, such as the pluton/country rock contact and along faults. As solidification of the magma proceeds, the outer crystallised shell of the pluton fractures by release of vapor pressure and this provides conduits for extraneous water infiltration into the intrusion. The resulting exchange between water and rock can produce large changes in the isotopic compositions of both materials if the isotopic gradient is large and the temperature favors rapid equilibration in the respective isotope systems.

The analysis of the ^{18}O composition of coexisting mineral phases in an igneous rock provides a way to tell

whether hydrothermal isotopic depletion has occurred during or after the magmatic stage. From experimental data and geological observation, the equilibrium $\delta^{18}O$ fractionation between quartz and feldspar in granitic rocks is estimated at about +2°/oo. This is an average value, since the measured $\Delta^{18}O$ of quartz-alkali feldspar pairs range from +0.7 to +1.6°/oo, while quartz-plagioclase pairs range from +1 to +2.5°/oo, depending on the anorthite content of the plagioclase (O'Neil & Taylor, 1967, Taylor, 1968, Blattner & Bird, 1974, Matsuhisa *et al.*, 1979). The great majority of low-^{18}O granitoids have $\Delta^{18}O$ (qtz-fp) larger than the equilibrium values reported above. A few granitoid rocks though, have mineral fractionations within the equilibrium range. It follows that for these rocks, the depletion of ^{18}O occurred during, or prior to the magmatic stage. Remelting of preexisting hydrothermally altered rocks or assimilation and exchange with altered wall rocks have been documented to produce low-^{18}O magmas (Muehlenbachs *et al.*, 1974, Hattori & Muehlenbachs, 1982).

Feldspars, of all major mineral phases in granites, are most likely to be affected by post-solidus hydrothermal exchange (O'Neil & Taylor, 1967, Taylor, 1974a, 1977). Quartz, conversely, is highly resistant to this type of alteration and tends to preserve its original ^{18}O composition if the temperature exchange is less than 400-500°C (Clayton *et al.*, 1972). The difference in the relative exchange rates of these two minerals is well

illustrated in figure 3-1. The diagram shows that for the study area, feldspar is depleted to a much greater extent than quartz by low-^{18}O fluids. The slope of the trend below the equilibrium line of $\Delta^{18}O$ (qtz-fp) = +2‰ is approximately four. This is in good agreement with a similar trend seen for coexisting quartz and feldspar in the Idaho batholith (Criss & Taylor, 1982). The latter authors postulated that the slope might be a direct reflection of the relative speed with which quartz and feldspar exchange ^{18}O in hydrothermal conditions.

Alternatively, diffusion and cation exchange experiments in laboratory conditions have shown that quartz is much slower in exchanging oxygen than the four to one slope seems to indicate. Diffusion experiments on albite (Mérigoux, 1968, Yund *et al.*, 1981, Giletti and Nagy, 1981) and quartz (Haul & Dümbgen, 1962) show that the rate of volume and grain boundary diffusion in feldspar is at least four to five orders of magnitude faster than quartz. Furthermore, the rate of oxygen exchange of feldspar in salt-rich solutions is even faster than observed in diffusion experiments since the oxygen exchange is obtained through cation exchange during solution-redeposition of the crystals (O'Neil & Taylor, 1967) and during hydrous mineral formation and albitisation (Cole & Ohmoto, 1976, 1979). Similar processes are difficult to ascribe to quartz oxygen exchange, since no recrystallisation or alteration is evident on the quartz grains. Intrinsic oxygen diffusion in

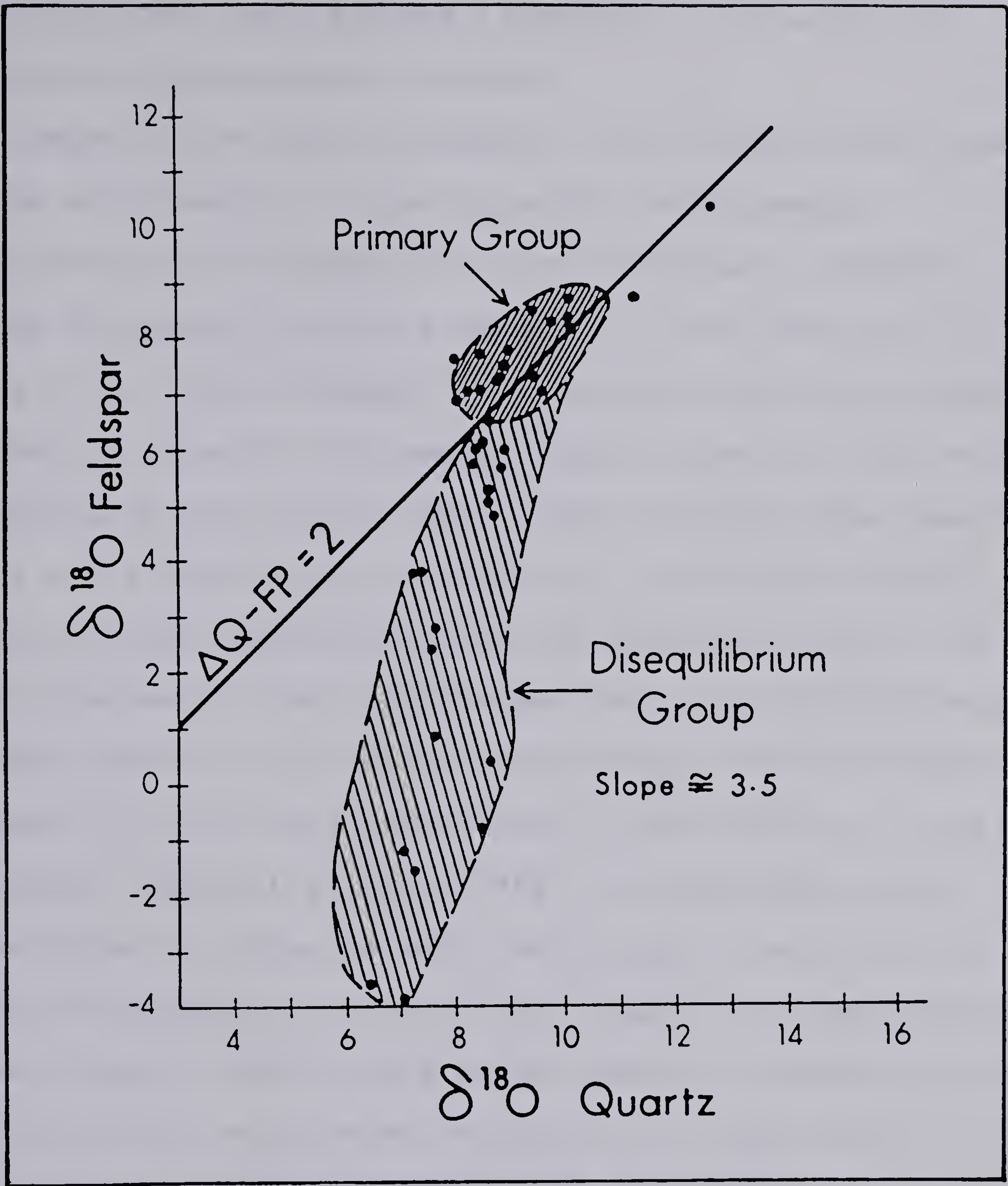


Figure 3-1 $\delta^{18}O$ of feldspar versus quartz. All available plutonic samples are plotted. The reference line Δ Q-FP is based on average fractionations observed in fresh granites (Taylor, 1968, 1974). The disequilibrium group below the reference line has a slope of about 3.5. The trend is virtually identical to the one observed by Criss and Taylor (1983).

quartz is the likely process affecting its rate of ^{18}O exchange in hydrothermal systems.

Based on the above arguments, the disequilibrium trend can be explained by (1) geologically instantaneous equilibration of feldspar with low-^{18}O meteoric fluids during alteration, with a time scale on the order of 100 years or so (Cole & Ohmoto, 1979) and (2) very slow oxygen diffusion in quartz for long periods of time (10^4-10^5 years) depending on the cooling rate of the intrusion. The observed slope would therefore be fortuitous; a reflection of the amount of time oxygen-diffusion had been operating in the quartz grains. In very shallow and fast-cooling hydrothermal systems, quartz would remain unexchanged, while in deeper systems, it would have enough time to equilibrate ^{18}O more thoroughly (Smith & Giletti, 1982). In this model, the coincidence of slopes between the isotopic composition of quartz and feldspar of this study (figure 3-1) and that of Criss & Taylor (1982) would be the result of broadly similar geological and hydrothermal conditions in both areas.

Other isotopic, geological and petrographic charateristics of low-^{18}O granites include:

1. Epizonal intrusion of magmas in highly jointed and fractured permeable country rocks.
2. Depletion of the ^{18}O of feldspars accompanied by petrographic evidence for the presence of alteration, such as clouding of the surface of the mineral grains and presence of saussurite.

3. Textural features such as micrographic quartz-alkali feldspar intergrowths and fine grain size associated with rapid cooling of magmas.
4. Ferromagnesian secondary mineral formation such as chlorite, uralite, epidote and oxides after hornblende, biotite and pyroxene.

All of the above features have been described in papers by H.P. Taylor (see reference list) and are seen in the samples of study area.

Temperature of hydrothermal fluids

The chl-ser-epi-ca † and subordinate sphene assemblage most commonly seen in the ^{18}O altered granitoid rocks can be characterised as part of the lower greenschist facies described in various metamorphic petrology textbooks (Mason, 1978, Winkler, 1979). An upper stability limit of 400°C under low pressure conditions (1-3 Kbar) is usually reported for the above assemblage in mafic igneous rocks. Indications are that the true limit is probably lower (approx. 350°C) in magnesian-poor, magnetite-rich acidic rocks.

Authigenic mineral groups, presently forming in contact with saline (4 molal) solutions of the Salton Sea geothermal system (McDowell & Elders, 1980) are nearly identical to those observed above. The Salton Sea metasedimentary rocks were originally fine grained feldspathic sandstones and interbedded mudstones and siltstones. Pre-metamorphic

----

† The abbreviations are as defined in the list provided at the beginning of the text.

assemblages consisted of clays, biotite , chlorite, muscovite, feldspar and quartz. The broad primary mineralogical similarities between these metasedimentary rocks and granites are obvious and suggest that the mineral succession described below can be used to estimate the temperature of hydrothermal cycling in the study area. The observed mineral paragenesis through the Salton Sea metasedimentary borehole section are summarised as follows:

1. At approximately 500 meters depth and 200°C, epidote first appears after carbonates and persists to the maximum depth drilled.
2. Matrix sericite is present at temperatures below 280°C. Flakes of sericite in feldspars persist throughout the calcite-chlorite zone (below) up to 325°C.
3. The calcite-chlorite zone spans temperatures of 190 to 325°C. Calcite disappears completely at the upper boundary, while chlorite persists as a minor phase up to 348°C.
4. Biotite appears as calcite disappears and persists to the maximum observed temperature of 360°C at a downhole hydrostatic pressure of 0.2 Kbar.
5. Andradite garnet appears at the bottom of the borehole while K-feldspar reacts out completely.

Objections might be raised about drawing parallels between geothermal alteration in a metasedimentary sequence such as the Salton Sea area and hydrothermal conditions near granitic plutons. But as the discussion that follows

suggests, the above temperature data is probably applicable to epizonal plutonic environments.

Geochemical studies on porphyry copper deposits throughout the western U.S.A. are in general agreement with the Salton Sea temperature results. The secondary mineral groups observed in the previous section, closely correlate to the propylitic alteration zone assemblages found at the periphery of porphyry deposits. Reported homogenization temperature of fluid inclusions coexisting with the mineral assemblages of qtz-kfp-chl-epi-py-mag and qtz-ser-py are 300°C for the Mineral Park, Arizona deposit (Wilkinson, 1981), 290 to 320°C for the Sierrita deposit (Beane & Titley, 1981) and 200-350°C for the New Mexico Santa Rita site.

The fluid inclusion temperature estimates agree with those inferred through oxygen isotope data (Sheppard *et al.*, 1971). The Santa Rita deposit yielded quartz-sericite isotopic temperatures in the range of 285 to 390°C, which is slightly higher than previously discussed, but generally within the same hydrothermal temperature conditions.

From the evidence above, one can reasonably conclude that for propylitically altered samples in batholiths, most of the chemical and isotopic exchange occurred between 200 and 350°C. This estimate will be used in the following sections to evaluate the isotopic composition of the fluids involved in the hydrothermal exchange.

A few samples described in the previous section do not follow the general rule of progressive propylitisation with increase in isotopic disequilibrium between minerals (WHA2, WHA8c and WHA10, among others). These samples typically do not contain hydrous minerals in appreciable amounts and the pristine appearance of the primary phases do not suggest the presence of major ^{18}O exchange. This type of association between disequilibrium $\delta^{18}O$ and fresh petrographic features has been observed mostly in rocks of quartz dioritic to dioritic composition. Taylor & Forester (1979) noticed this type of relationship in the gabbroic Skaergaard intrusion and concluded that temperatures involved in the hydrothermal event were higher (400-600°C) than those encountered in more acidic rocks. These temperatures are beyond the normal stability of hydrous minerals such as chlorite, serpentine and clay minerals and this is the reason only trace amounts of secondary hornblende are seen in the late-stage granophyres of the Skaergaard. A study of heat and mass flow through that intrusive (Norton & Taylor, 1979) confirms this. It shows that 75% of all fluids involved in hydrothermal cycling, were hotter than 480°C. The samples of this study are more siliceous than the Skaergaard intrusion and so their solidus temperature must have been lower (≃1000°C for the Skaergaard [Norton & Taylor, 1979] and ≃650-750°C for granodiorites [Merrill *et al.*, 1970]). Therefore, the initial temperature of hydrothermal circulation in these intrusions should also be lower than

the Skaergaard temperature estimates. Although the precise range of alteration temperatures is not known for the quartz diorites of this study, temperatures greater than about 400°C are required in order to impede hydrous mineral formation.

B. The Whitehorse Trough area - a case study

The Cretaceous-Tertiary intrusives in the Mesozoic volcanic sedimentary rocks of the Whitehorse area provide an ideal case to study the effects of hydrothermal alteration involving meteoric water. It is the most densely sampled area of this study and permits estimation of the volume of fluids involved and the hydrothermal conditions that existed during the period of ^{18}O exchange. Statistically, the most pervasive alteration occurred in this area. Figure 3-2 shows that of the 9 plutonic bodies studied, six show disequilibrium $\delta^{18}O$ fractionations between minerals as well as some examples of isotopic reversals. Along with feldspars, biotite and magnetite show depleted $\delta^{18}O$ values. This association is to be expected since most altered rocks have ferromagnesian minerals rimmed with secondary magnetite, that formed contemporaneously with epidote and chlorite.

The immediate area surrounding the city of Whitehorse has most of the highly ^{18}O-exchanged rocks of the Trough (figure 3-3, Table 2-1). They include the Cap Mountain suites, the Whitehorse batholith with its associated copper

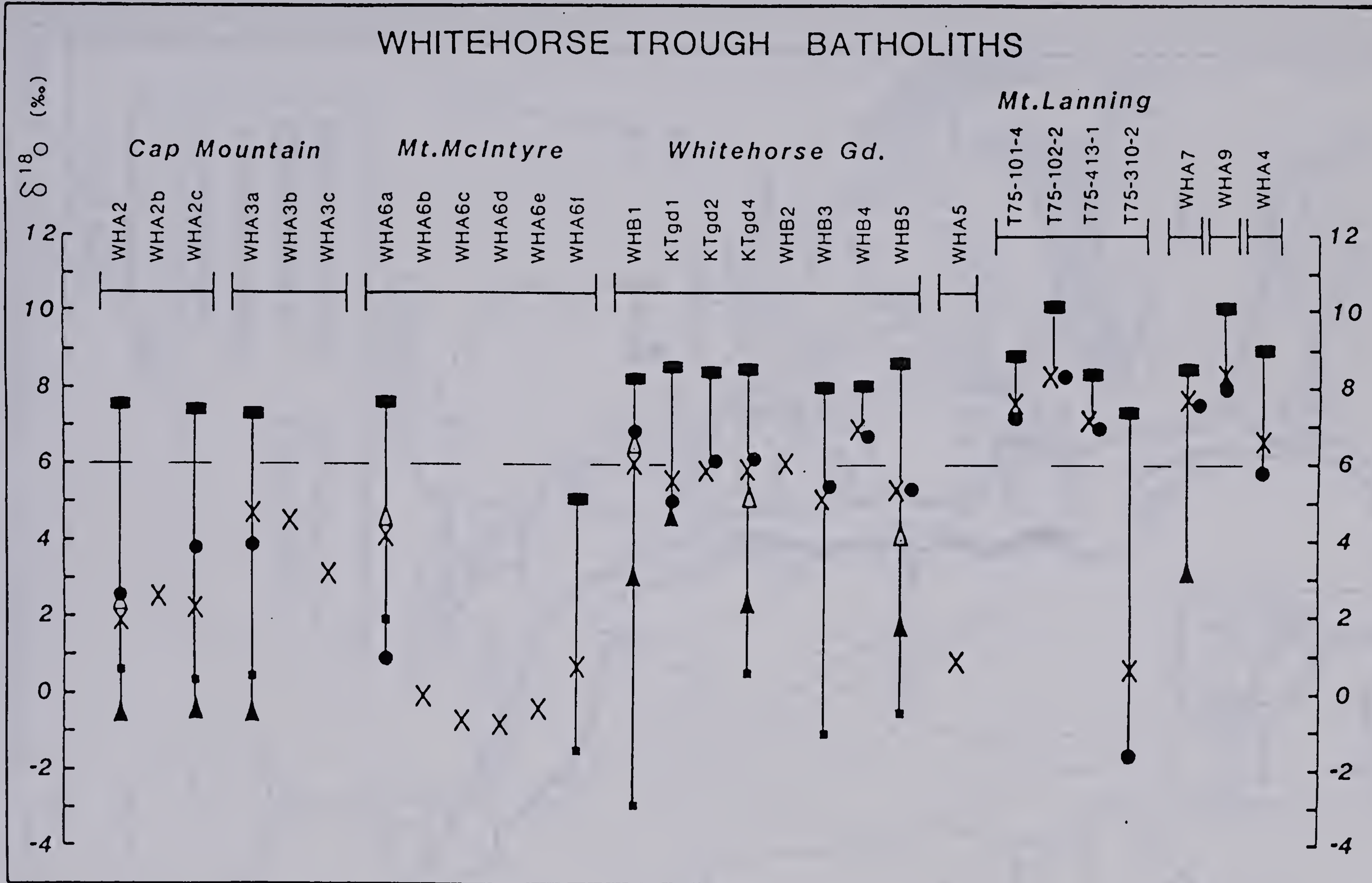


Figure 3-2 $\delta^{18}O$ results for the Whitehorse Trough granitoid batholiths. Rectangles → quartz, dots → feldspar, X → whole-rock, filled triangles → biotite, open triangles → hornblende, and squares → magnetite. Values below the +6 limit (dashed line) are thought to represent altered samples.

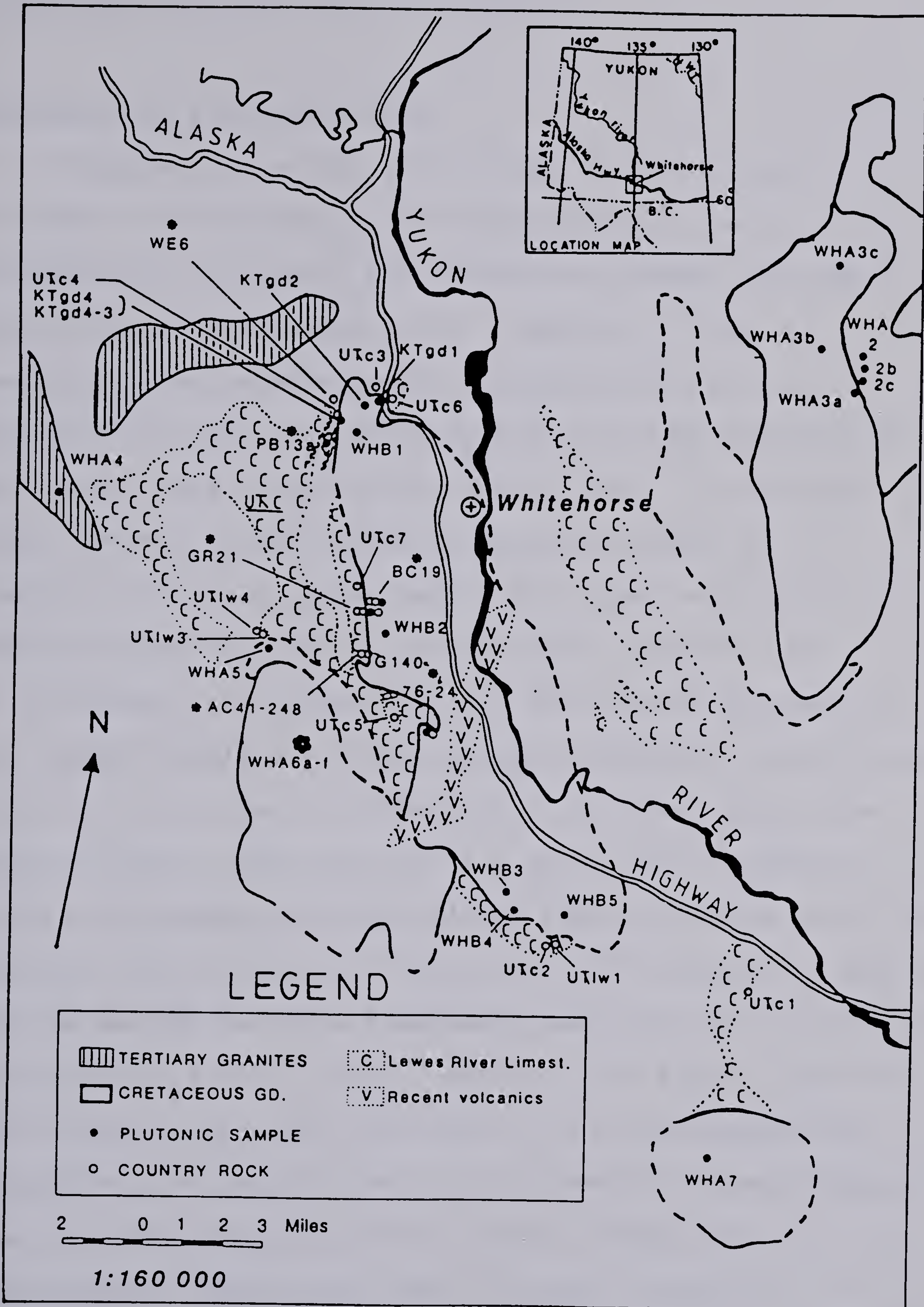


Figure 3-3 Location of samples in the Whitehorse area. Symbols are explained in the legend. Asterisks show G. Morrison's rocks.

skarns and the Mount McIntyre batholith.

Cap Mountain granitoid rocks

Cropping out on the north ridge of Cap Mountain, northeast of Whitehorse, two mineralogically (and isotopically) distinct rock types were studied. The WHA 2 group forms an elongated pluton composed of fine- to medium-grained hornblende-biotite quartz diorite to granodiorite which was dated by K-Ar at 92 Ma (Morrison *et al.*, 1979). The second suite, WHA 3 a to c, is possibly older (98 Ma?) than the WHA 2 rocks and ranges in composition from a coarse leucocratic granite porphyry to a medium-grained biotite quartz monzonite. Despite the mineralogical and whole-rock $\delta^{18}O$ differences between the two suites (table 2-1) their mineral $\delta^{18}O$'s are surprisingly similar and systematic (figure 3-2). At first sight, the pattern seems compatible with the proximity of the two suites and suggests similar fluids were in contact with both plutons. The calculated † whole-rock $\delta^{18}O$ values for samples WHA 2a and WHA 3a agree reasonably well with their measured $\delta^{18}O$s (2.4 & 4.5°/₀₀ [calc.] versus 1.9 & 4.7°/₀₀ [meas.]). This implies that that the final $\delta^{18}O$ of the samples was determined entirely by the relative amounts of each phase in the rock. The picture is more complex though, as petrographic observations show. The WHA 3 group is pervasively altered to clay minerals and chlorite with minor

---

† The calculation is based on modal abundances shown in appendix 1.

sericite while the more depleted suite, WHA 2, shows relatively little secondary alteration (figures 3-4 & 3-5). Saussurite in the calcic cores of zoned plagioclase crystals does occur in patches of both groups but to a lesser extent in WHA 2. Pristine hornblende is found in both intrusives but clear unchloritised biotite is only seen in the lower ^{18}O WHA 2 group.

A higher temperature of alteration, which was explained previously in this chapter, is a possible explanation for the absence of propylitisation in the WHA 2 suite. Upon further analysis though this does not seem to provide an entirely satisfactory explanation for the isotopic relationships shown in figure 3-2.

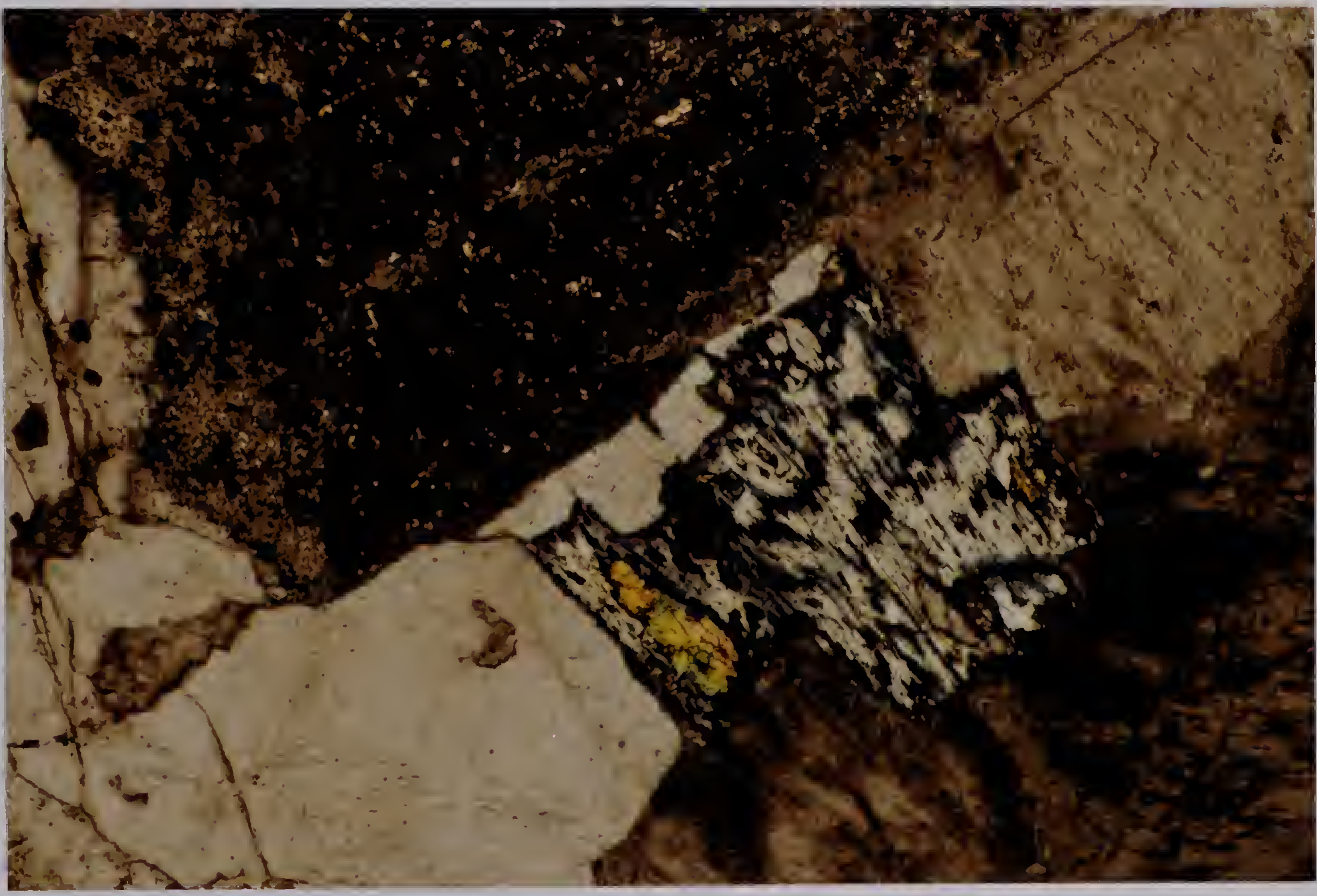
At higher temperatures, all other parameters being constant, rates of exchange increase and fractionations between the various minerals and the hydrothermal solutions decrease. This would require that the final ^{18}O composition of the fluids in contact with the respective suites be significantly different; $\Delta^{18}O$ fluid $_{2-3}$ = 1-2‰ at T_2 = 400-500°C and T_3 = 250-300°C) †, assuming that $\delta^{18}O$ feldspar ≅ $\delta^{18}O$ rock and that final composition of the rock was in equilibrium with the hydrothermal fluids. Given this and the fact that minerals have distinct rates of isotopic exchange that vary with temperature, it is highly unlikely that the final $\delta^{18}O$ of quartz, biotite, feldspar and magnetite in

----

† The subscripts 2 and 3 refer to suites WHA 2 and WHA 3. Matsuhisa *et al.*'s, 1979, feldspar-water fractionation equation was used in the calculation.

Figure 3-4 Thin section photograph of sample WHA 2c (crossed nicols). The magnification is X 40. The rock is an alteration-free granodiorite-quartz diorite yet is depleted in ^{18}O down to +2.2 permil. Field of view spans 2.6 mm.

Figure 3-5 Thin section of WHA 3a in crossed nicols. (X 40). Upper dark mineral is fully saussuritised plagioclase feldspar crystal. Central mica mineral is chlorite with inclusions of epidote. The lower right-hand side dark mineral is an altered alkali feldspar. The shading is the result of clay minerals on its surface. White mineral is quartz.



both the higher and lower temperature suites would eventually be equal. Unfortunately, the isotopic data is insufficient to exactly define the geological processes that gave rise to such similar mineral $\delta^{18}O$ values in both Cap Mountain suites. Microprobe, X-ray diffraction, D/H and fluid inclusion studies could probably resolve this problem by identifying more thoroughly the mechanism and the extent of alteration, the chemistry of the phases, the exact nature of altering fluids and the temperatures at which waters circulated. Future investigations by these methods should prove to be a fruitful avenue of research in difficult cases such as this.

The Whitehorse batholith

Because of its association with economically viable skarn deposits and its ease of access, the Whitehorse batholith has been sampled most extensively along with the limestone strata surrounding it (figure 3-3). According to Morrison (1981), the batholith is a single synorogenic pluton of slightly alkali-enriched calc-alkaline character. The western contact with carbonate rocks is steep, locally overhanging, with many apophyses and embayments occurring near the skarn ores. The northern contact is shallow and outward-dipping while the eastern part of the intrusion is buried beneath Quaternary sediments. Its composition varies gradationally from a core of biotite quartz monzonite to a grey hornblende-biotite quartz diorite at the western

margin. This variation is thought to be the result of progressive admixture of microxenoliths of calcic plagioclase (An_{60} and augite (partly to totally transformed to hornblende) from a basic volcanic source, most likely the Upper Triassic Lewes River basalt, into a magma of quartz monzonitic composition. These microxenoliths are partially disaggregated and preferentially altered relative to the more felsic host intrusion. Large xenoliths up to 60 cm wide of varying composition are found sporadically throughout the batholith but locally have been observed to comprise 60% of the volume of the quartz diorite (Morrison, 1981).

One inclusion sample (KTgd-4-3), collected by the author north of the Pueblo Mine has been analysed along with pyritic siltstone from which it came (uTc-4).

Results of the granodiorite sample measurements (figure 3-2) and calc-silicate rocks, uTc-3, -4 and KTgd-4-3 (+5.9, +6.1, +4.0‰) close to the pluton contact (30 meters to 1 cm) show a very high degree of homogeneity of the whole rock $\delta^{18}O$ values. The $^{18}O/^{16}O$ ratios of quartz, feldspars and hornblende in this suite also show little variation among the samples. Biotites are the most consistently altered and exhibit a much larger spread in $\delta^{18}O$ (4.5 to 1.7‰). This is a reflection of the relative amounts of chlorite present in biotite as evidenced by its low K contents (Morrison *et al.*, 1979, appendix 1A). A similar spread (-3 to +0.4‰ is seen for magnetite which is closely associated with the breakdown of biotite.

Samples collected by G. Morrison and analysed at the University of Western Ontario stable isotope laboratory (appendix 3) are, on average, 2.2°/₀₀ enriched in ^{18}O relative to our results. Only whole-rock data is presented in the former study (Morrison, 1981) so is not possible to ascertain the relative importance of disequilibrium processes in these rocks, if they exist. Morrison's samples were all collected near contact alteration zones, mostly within 2 meters of the skarn contact. Except for Ktgd-4, our granitic samples were at least 25 meters away from the surrounding country rocks and most were well within the margins of the intrusion. Our only contact samples (figure 3-3) from the northern part of the granitic body, are not strictly analogous to Morrison's rocks since (1) no significant mineralisation was present at the site and (2) the country rocks were not metamorphosed carbonates but were part of the pyritic siltstone unit.

In view of their location (near carbonate units) and the narrow $\delta^{18}O$ range for all our batholith samples, Morrison's samples are not thought to represent the original composition of the Whitehorse granodiorite. They probably became enriched in ^{18}O by exchange with the immediately adjacent limestones (average primary $\delta^{18}O$ = +24°/₀₀). This could have been done partly by diffusion of wall rock ^{18}O into the magma prior to the hydrothermal event that produced the mineralised skarns. A much faster and more likely process involves interaction of ^{18}O enriched CO_2 produced

during contact metamorphic decarbonation of the Lewes River limestones such as has been observed at other similar geological sites (Deines & Gold, 1969, Shieh & Taylor, 1969b and B.Taylor & O'Neil, 1977).

At the estimated temperatures of contact metamorphic alteration (550+°C, see Morrison, 1981), the $\delta^{18}O$ of CO_2 gas in equilibrium with the average $\delta^{18}O$ of the limestones would be about +30°/₀₀ (Bottinga, 1968, O'Neil & Epstein, 1966b) based on experiments and theoretical calculations. The CO_2 gas, mixing with either magmatic water or H_2O derived from the dehydration of the interbedded argillaceous units, could increase the total $\delta^{18}O$ of the fluid by 2.2°/₀₀, if the mole fraction of CO_2 in water was equal to 0.1. Furthermore, isotopic exchange, at these temperatures, between the CO_2 gas and the H_2O components of the fluid would increase the $\delta^{18}O$ of the water by about 1°/₀₀ (B.Taylor & O'Neil, 1977). Through this process of ^{18}O enrichment, metamorphic or magmatic waters near the contact zone would obtain much higher $\delta^{18}O$ values than the Whitehorse pluton. Any high temperature (400-500°C) isotopic interaction of these fluids with the intrusion would most likely produce the isotopic enrichment observed at its margins.

Most orebodies in the Whitehorse Copper Belt are located in skarn pendants surrounded by the granodiorite or in embayments along its margins. Mineralisation was therefore concentrated around so-called heat-islands close to which hydrothermal activity was enhanced. If

decarbonation and dehydration of the pelites occurred prior to the meteoric phase then the fluid produced would undoubtedly be high in ^{18}O (+15 to 20°/oo). Because of the geometry of the mineralised pendants (cf. Tenney, 1981, Morrison, 1981, figs.16-26) such fluids must have come in contact with and perhaps percolated through the margins of the batholith. The temperatures of skarn formation being high (approx. 500°C), secondary hydrous mineral formation in the altered granodiorite margin need not occur.

Previous stable isotope studies by Shieh & Taylor (1969a, b) and Turi & Taylor (1971) on contact interaction effects have shown that pluton margins are very susceptible to ^{18}O exchange with the surrounding country rocks. These isotopic effects are especially visible when the isotopic gradient between rock types is large. Nagy and Parmentier (1982) showed that oxygen diffusion through a fluid phase could account for the ^{18}O increase seen within 60 cm of the contact of the Sawtooth Stock with surrounding metasedimentary rocks (Shieh & Taylor, 1969a). The $\delta^{18}O$ values of the Osgood Mountain granodiorite increase by 2°/oo within 2 meters of the contact with carbonates and skarns (B. Taylor & O'Neil, 1977), a situation and an ^{18}O enrichment very similar to the one observed in this study.

Finally, copper concentrations determined by Morrison show a variation from the barren core (19 ppm) to the contact altered margins (91 ppm). The similarity of the latter concentration with copper measurements of the

silstone units (55 ppm) and the Lewes River volcanic rocks (95 ppm) support a country rock-batholith margin exchange hypothesis.

Equilibrium versus disequilibrium in the Whitehorse granodiorite

As was pointed out previously, a 2‰ fractionation between quartz and feldspar is considered normal for granitic rocks. It is useful to separate those rocks that have suffered meteoric ^{18}O exchange from pristine samples. In many cases though, "wet" acid plutonic rocks exhibit retrograde ^{18}O exchange without major involvement of depleted meteoric water (Bottinga & Javoy, 1975, Javoy, 1977, Deines, 1977, Matsuhisa *et al.*, 1979). Such retrogression typically involves feldspar and hydrous minerals such as biotite and hornblende and is especially prominent in pegmatitic rocks (Longstaffe *et al.*, 1981, Longstaffe, 1982). Isotopic temperatures based on experiments and theoretically-derived fractionation curves were calculated for the rocks of the Whitehorse area (Table 3-1). The rather wide calculated temperature range found within intrusives indicates that minerals are in isotopic disequilibrium with each other and that the temperatures are essentially geologically meaningless. They reflect the different rates of exchange at subsolidus temperature of the minerals in the rock.

Table 3-1 : Isotopic temperatures from the Whitehorse area granitoid rocks.

Sample	Mineral Pairs: Qtz-Bio [a]	Qtz-Hbl [a]	Qtz-Mag [b] [a] [c]	Qtz-Fpl [d] [e]	Bio-Mag [a]
WHA 6a		675 (3.2)	680-707-773 (5.8)	162-103 (6.8)	
WHA 6f			620-646-671 (6.6)		
WHA 2a	374 (8.2)	476 (5.3)	595-619-630 (7.0)	224-154 (5.2)	(-1.2)
WHA 2c	386 (7.9)		590-613-621 (7.1)	324-233 (3.6)	(-0.8)
WHA 3a	386 (7.9)		600-625-640 (6.9)	333-240 (3.5)	(-0.8)
WHB 1	530 (5.2)	923 (1.9)	400-429-405 (11.3)	623-414 (1.6)	257 (6.1)
WHB 3			490-505-479 (9.2)	404-294 (2.8)	
WHB 5	428 (6.9)	545 (4.4)	490-509-483 (9.1)	351-254 (3.3)	546 (2.2)
KTgd-1	613 (4.1)			324-233 (3.6)	
Ktgd-4	465 (6.2)	582 (4.0)	540-561-548 (8.0)	454-334 (2.4)	612 (1.8)
WHA 7	517 (5.3)			763-486 (1.0)	

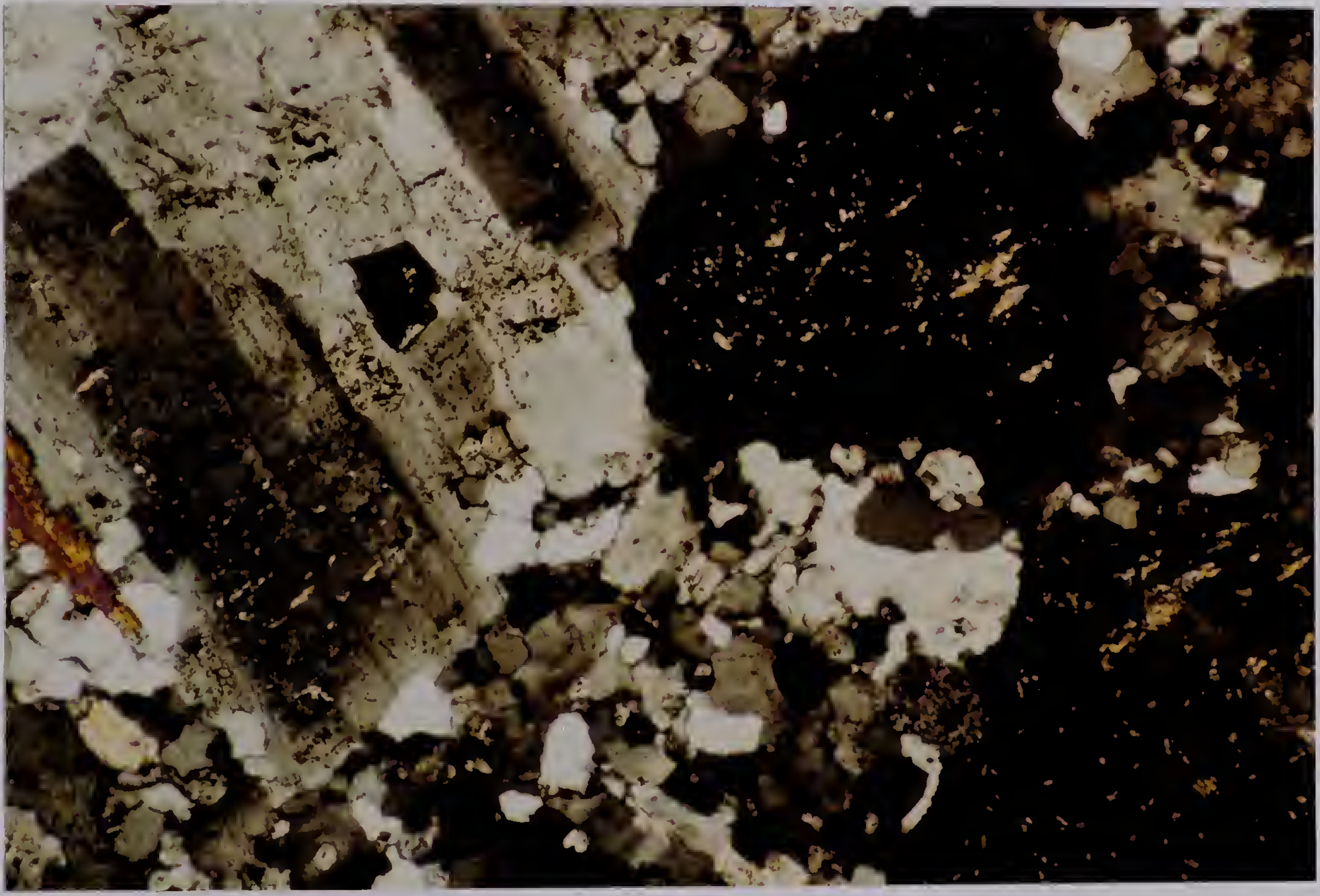
All temperatures are in degrees Celsius. Values in parentheses below the temperature values represent the mineral-pair fractionation in permil. Bracketed letters refer to the source of the isotopic partitioning equations, for the respective mineral pairs, used in calculating the above temperatures:
[a] Bottinga & Javoy (1975)
[b] Friedman & O'Neil (1977)
[c] Anderson et al (1971) and Matsuhisa et al (1979)
[d] O'Neil & Taylor (1967)
[e] Matsuhisa et al (1979)

Alteration in the Whitehorse batholith samples is slight to moderate and occurs in patches (figures 3-6, 3-7). Morrison suggested that it was the result of the chemical disequilibrium between the calcic phenocrystic phase and the more acid matrix. Unlike the other plutons in the area, meteoric water is not necessary to explain the isotopic signature of the Whitehorse batholith. The small spread in whole-rock $\delta^{18}O$ and the normal and near-normal quartz-feldspar fractionations suggest retrograding of feldspar and biotite in a wet environment where the fluid was approximately in equilibrium with the intrusion. This conclusion is supported by sample WHB 4 ($\delta^{18}O = +6.9°/_{oo}$) which is the most mineralogically altered rock of this suite, yet shows the smallest quartz-feldspar $\delta^{18}O$ fractionation ($1.3°/_{oo}$). This indicates that even though local alteration did occur, the $\delta^{18}O$ of the fluid was not significantly depleted. If this is true, then the Whitehorse batholith must have been a low-^{18}O magma of +5 to $+6°/_{oo}$ composition during or just after emplacement.

Primary low-^{18}O granites ($\delta^{18}O = -0.6$ to $+2°/_{oo}$) have been inferred in Scotland's Tertiary igneous province (Forester & Taylor, 1977) and in Oregon (Taylor, 1971). In both cases, the plutons were part of the latest phase of igneous activity in a region having suffered a long history of plutonism and volcanism. As such, the incorporation of hydrothermally altered and depleted roof rocks into the magma chambers of ascending intrusives, or wall-rock - melt

Figure 3-6 WHB 1 in crossed nicols (X 31). Thin section shows patchy alteration of plagioclase (left-hand side) and hornblende (right-hand side). Alteration products include sericite and epidote. The central plagioclase and quartz are unaltered. Field of view spans 3.3 mm.

Figure 3-7 <u>WHB 4 in crossed nicols (X 31)</u>. It shows patchy alteration also. The sample is more heavily altered than any other WHB rock. The two dark minerals are saussuritised plagioclase cores. The matrix is less intensely altered to clay minerals.



^{18}O exchange are plausible explanations for the low $\delta^{18}O$ values encountered in these regions In the case of the Whitehorse batholith, a similar cause for its low-^{18}O character is not compatible with the timing of the emplacement of the pluton. K-Ar mineral dates and Rb-Sr isochrons (Morrison *et al.*, 1979) show that the pluton predates all the other granitoid rocks sampled in the area and therefore low-^{18}O country rocks were not likely to be present at that time ($\simeq$ 115 Ma ago).

An alternative explanation is suggested by G. Morrison's assertion that the Whitehorse magma was strongly convecting and assimilated Lewes River-type basaltic material during its emplacement. Unaltered island-arc andesites and basalts typically have $\delta^{18}O$ values ranging from +5.5 to +6.5‰ (Taylor, 1968, Matsuhisa *et al.*, 1973, Matsuhisa, 1979). If enough basaltic material of $\delta^{18}O$ of +5.5‰ was incorporated into a qtz monzonitic magma of approximate +7.5‰ composition, then it is possible that a granodioritic magma of intermediate ^{18}O composition +6 to +6.5‰) would result. Mass balance considerations though, indicate that equal amounts of basaltic and plutonic material would be required to produce the final $\delta^{18}O$ value above. This is a much larger mass of country rock than can be incorporated into the entire volume of a cooling pluton, as was discussed by Taylor (1980). This process could only have been sufficient in scale, locally, in the roof zone of the batholith or near its wall margins. The compositional

gradation from a quartz dioritic margin to a quartz monzonitic core, in the batholith, is compatible with the constraints above. Unfortunately, no truly silica-rich samples were collected or analysed from this intrusion, and the final test of the model is still to be made.

Country rocks

As opposed to the relatively uniform $\delta^{18}O$ values inside the batholith, faulting and shearing during plutonic emplacement (Morrison, 1981, figs.16 to 26) and the associated hydrothermal activity has produced variable but generally low-^{18}O compositions in the country rocks near the intrusion. The skarn samples from Morrison's thesis (ref. cited) and some country rocks from this study (Table 3-2 and figure 3-3) show evidence of involvement of meteoric water in the isotopic exchange process. Six diopside skarn samples range from -9.8 to +6°/₀₀. Most limestones and marbles from the immediate area are consistent with a marine environment of deposition, averaging a $\delta^{18}O$ of +24°/₀₀ and a $\delta^{13}C$ of +0.1°/₀₀. The Precambrian marbles from other regions (figure 3-8) are poorer in ^{18}O by 3 to 6 °/₀₀. This is to be expected since carbonates and cherts in many areas of the world have shown a general increase in ^{18}O throughout geological time (Degens and Epstein, 1962, Knauth & Lowe, 1978, Perry *et al.*, 1978). Only one sample of Lewes River carbonate (uTc-7) is consistent with a meteoric-hydrothermal process (*i.e.* a depletion in ^{18}O without much change in the

Table 3-2 : Country rock $\delta^{18}O$ and $\delta^{13}C$ results

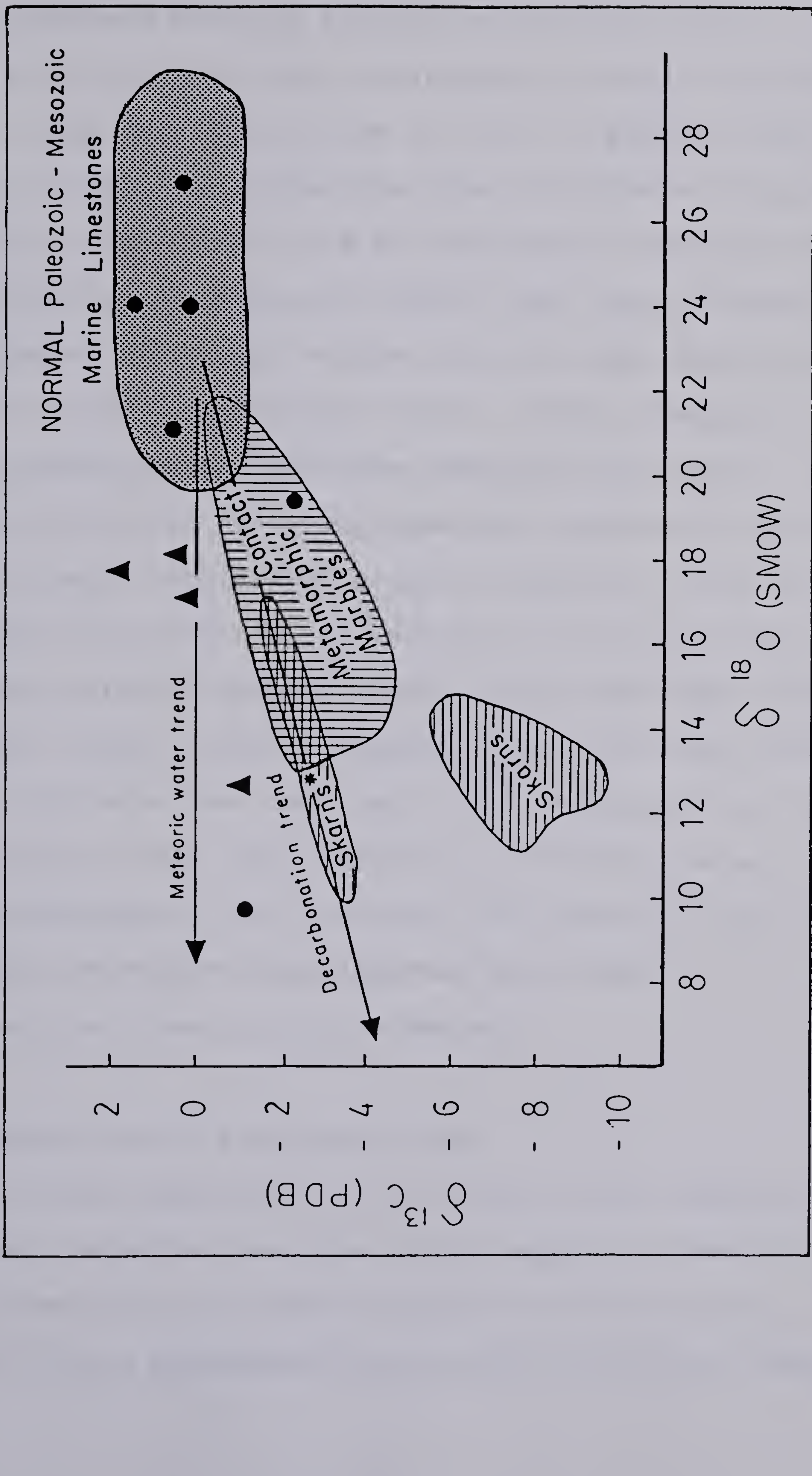
Sample	Rock Type	$\delta^{18}O$ ‰	$\delta^{13}C$ ‰
YUKON CRYSTALLINE TERRANE			
HCsn-1	Precambrian biotite schist	7.8	
Hc-3	Hadrynian coarse white marble	18.1	0.6
PPsn-1	Paleozoic musco-biotite schist	7.0	
COAST PLUTONIC COMPLEX			
JKd-1	Jurassic? greywacke (Dezadeash Gr)	10.2	
JKk-1	Jurassic? biotite schist (Kluane)	13.0	
JKk-2	Greyish tectonised slate (Kluane)	12.7	
JKk-3	Biotite schist,micaceous quartzite	10.4	
JKk-4	Precambrian greenstone (Yukon Gr)	-7.9	
OMINICA CRYSTALLINE BELT			
Hc-1	Hadrynian dark fine limestone	12.7	-1.1
Hc-2	Hadrynian dolomitic marble	17.1	0.4
lCc-1	Lower Cambrian grey fine marble	17.6	2.0
HPm-1	Paleozoic (or older) bio schist	9.6	
CPsn-1	Paleozoic hbl-sericite schist	11.4	
CPsn-3	Paleozoic chl-sericite schist	9.5	
CPsn-4	Paleozoic schistose greenstone	8.9	
CPsn-5	Paleozoic bio-chlorite schist	7.6	
DMs-2	Devonian magnetite-bearing slate	12.7	
CPv-4	Late Paleozoic fine andesite?	5.0	
Tv-2	Triassic sheared metavolcanic	11.3	

Table 3-2 (continued)

Sample	Rock Type	$\delta^{18}O$ ‰	$\delta^{13}C$ ‰
WHITEHORSE TROUGH			
CPv-1	Late Palezoic greenstone(Taku Gr)	9.1	
CPv-2	Late Palezoic serpentinite (Taku)	-1.1	
Tv-1	Triassic dacitic? tuff	11.2	
uTlw-1	Dioritic fine grained dyke	7.9	
uTlw-3	Late Triassic mafic hornfels (LR)	-0.6	
uTlw-4	Late Triassic metavolcanic (LR)	-1.1	
uTlv-1	L.Triassic andesite (Lewes River)	-0.8	
uTlv-2	L.Triassic olivine basalt (LR)	7.5	
uTlv-4	L.Triassic argillite/siltstone	11.3	
uTc-3	Diopsidic (malachite) skarn (LR)	5.9	
uTc-4	Diopsidic calc-silicate rock	6.1	
KTgd-4-3	Inclusion of uTc-4 in granodiorite	4.0	
Jl-1	Jurassic Siliceous Hornfels (LG)	7.5	
Jl-3	Jurassic arkosic sandstone (LG)	14.1	
uTc-1	Grey fine limestone (Lewes River)	26.8	0.6
uTc-2	Black fine grained marble (LR)	23.9	-0.7
uTc-5	Greyish-Black fine limestone (LR)	21.0	-0.4
uTc-6	Greyish fine carbonate siltstone	24.0	0.7
uTc-7	Malachite stained medium marble	9.7	-1.1
uTlv-3	Greyish coarse marble (LR)	19.3	-2.2

All $\delta^{18}O$ values are in permil (SMOW), $\delta^{13}C$ results - (PDB).
() represent the names of formations or rock groups.

Figure 3-8 <u>$\delta^{13}C$ - $\delta^{18}O$ plot of limestone samples.</u> Circles are Lewes River limestones; triangles are Precambrian marbles and limestones. The dark shaded area is the normal range for marine carbonates. The horizontal trend for meteoric water alteration does not affect the ^{13}C composition of carbonates because water has no intrinsic CO_2 reservoir. The decorbonation trend trend arises during fractional distillation of isopically enriched CO_2 from the residual calcite. Contact metamorphic marbles and calc-silicate rocks follow this trend approximately (Deines & Gold, 1969) The contact altered metacarbonates and skarns of Shieh & Taylor, 1969b also plot around the decarbonation line (skarns*). The lower skarn region is from B. Taylor & O'Neil (1977). These rocks may have been depleted through oxidation of carbon during fluid circulation.


NORMAL Paleozoic - Mesozoic
Marine Limestones
Contact
Metamorphic
Marbles
Skarns
Skarns
Meteoric water trend
Decarbonation trend
$\delta^{13}C$ (PDB)
2
0
-2
-4
-6
-8
-10
$\delta^{18}O$ (SMOW)
8
10
12
14
16
18
20
22
24
26
28

^{13}C content of the carbonate). Contact metamorphic decarbonation is a plausible scenario for uTlv-3, a marble near the Montana Mountain batholith (Carcross area). Residual calcite left after decarbonation should follow a line of slope of 0.2 to 0.3 on a $\delta^{18}O$-$\delta^{13}C$ plot, if the process is similar to a Raleigh-type distillation (figure 3-8). This trend is followed by the Trenton limestone near the Mt-Royal pluton (Deines & Gold, 1969) and by contact metamorphosed skarns and marbles near the Beer Creek quartz monzonite in Arizona (Shieh & Taylor, 1969b). Samples falling significantly below the decarbonation trend (B.Taylor & O'Neil, 1977) may have been depleted by low-^{13}C fluids through oxidation of graphitic material. None of the limestones of this study are amenable to such a process since their organic carbon content is low (Morrison, 1981). The normal oxygen isotopic composition of the Lewes River marbles indicates that they were quite impermeable to fluid flow except at very close proximity to the faulted and mineralised zones of the batholith. The skarns in such an environment must have been depleted by as much as 30‰ by large amounts of meteoric groundwaters.

$\delta^{18}O$ composition of altering fluids

The large depletion in ^{18}O of the skarns, carbonate rocks and greywackes near the pluton require a low-^{18}O fluid. Identifying the exact composition of the waters involved in the hydrothermal alteration is difficult because

of isotopic disequilibrium between minerals (above). For such purposes, vein minerals are the best material to analyse since they precipitate and grow directly from the hydrothermal fluids. Morrison has reported $^{18}O/^{16}O$ ratios of quartz-filled cavities and a quartz-vein (appendix 3). Using his estimates of temperatures at those localities (550 & 480 °C) and experimentally-derived quartz-water fractionation curve of Matsuhisa and others (1979), yields water with an ^{18}O composition of about +1.5 to 2°/oo for the garnet skarn alteration stage and -16°/oo for the late hydrous veining stage. The former value, which is 8°/oo richer in ^{18}O than the host skarns, indicates that the water from the cavities must have exchanged isotopically with high ^{18}O carbonate rocks from outside the low-^{18}O skarn aureole. Another possible explanation is mixing of fluids emanating from the batholith (+6 to +8°/oo) with depleted water in the skarn contact.

The late stage water (-16°/oo) is at least 5°/oo enriched compared to the well water sample (-21.4°/oo) from the Dezadeash Lodge and to meteoric waters in the Yukon-N.W.T. southern border area (-23.1°/oo, from Bowman & Covent, 1983) The former is the same as Magaritz & Taylor's (1976a, b) calculated $\delta^{18}O$ value for the alteration fluids of the Coast Plutonic rocks and the Intermontane Belt to the south. In view of the very low $\delta^{18}O$ values of the country rocks and granites, the -16°/oo water, at least as a starting point, is thought to be a realistic estimate for

meteoric water in Cretaceous-Tertiary time in the Whitehorse Trough. The skarn formation and the associated mineralisation was definitely a meteoric water-dominant system.

Water-rock ratios

Precise estimates of the amounts of water cycling through the batholith are difficult for the same reasons as outlined in the previous section. An added variable is whether the system was closed or open to continuous recharge with unexchanged meteoric water from above. By using realistic end-member ^{18}O compositions, minimum ratios of water to rock can be calculated. The molar ratio of water oxygen to rock oxygen in a closed system exchange intergrated over its history can be estimated by;

$$W/R = \frac{\delta^{18}O_{rock}(f) - \delta^{18}O_{rock}(i)}{\delta^{18}OH_2O(i) - \delta^{18}OH_2O(f)}$$

where *f* and *i* refer to the final and initial isotopic compositions of the materials. The final $\delta^{18}O$ of the exchanged water can be calculated assuming that the whole-rock $\delta^{18}O$ is approximately equal to that of plagioclase of appropriate anorthite content. Obviously the calculation is based on the premise of achievement of isotopic equilibrium between the rock and the final composition of the water.

Table 3-3 shows the calculated ratios obtained for a closed hydrothermal system in the Whitehorse area. As

Table 3-3 : Water-rock ratios in the Whitehorse area

Rock Suite	Average $\delta^{18}O$ °/oo	Temp. °C	W/R [A]	W/R [B]
WHA 6a-f	0.5	200	0.89	0.73
		300	0.60	0.56
		400	0.50	0.49
WHA 2a-c	2.2	200	0.55	0.47
		300	0.40	0.37
		400	0.34	0.33
		500	0.31	0.31
WHA 3a-c	4.1	200	0.30	0.26
		300	0.22	0.21
		400	0.19	0.19
WHB 1-5	5.8	200	0.13	0.11
		300	0.10	0.09
		400	0.09	0.09
WHA 4	6.6	200	0.06	0.06
		300	0.05	0.05
		400	0.05	0.04

$\delta^{18}O$ rock =+7.5°/oo, $\delta^{18}O$ H_2O =-16°/oo
Water-rock ratios were calculated assuming a closed hydrothermal system. The final $\delta^{18}O$ composition of the rock was assumed to be in isotopic equilibrium with hydrothermal water. Also to calculate the final $\delta^{18}O$ value of the water, we assumed that $\delta^{18}O$ rock = $\delta^{18}O$ feldspar of appropriate An content (30).
[A] = $10^3 \ln\alpha$ (Fp-water) = 2.68 $10^6 T^{-2}$ - 3.37 (O'Neil & Taylor,1967)
[B] = $10^3 \ln\alpha$ (Fp-water) = 2.12 $10^6 T^{-2}$ - 2.6 (Matsuhisa et al,1979)

starting parameters, a value of +7.5°/₀₀ was chosen for the initial $\delta^{18}O$ of the granitoids, based on averages of unaltered rocks from the Coast Plutonic Belt and the Whitehorse Trough (Chapter 2). The starting water composition was -16°/₀₀ as explained above while the anorthite content was estimated at 30 mole percent for the granodiorites in the area. The water-rock ratios do not depend very much on the choice of feldspar-water fractionation curves. They range from 0.05 to 0.9, a variation nearly identical to the one calculated for the Topley intrusions of Early Cretaceous age in the Intermontane Belt near Prince George, B.C. (Magaritz & Taylor, 1976b). Somewhat smaller values are expected if the hydrothermal system was open to some recharge with unexchanged groundwaters from above, as is likely in epizonal settings (Taylor, 1974a, 1977). The large water/rock estimates are common in volcaniclastic terranes in other areas of North America and Scotland (Taylor, 1978, Forester & Taylor, 1977) and are not surprising considering the permeabilities of the country rocks.

The lower W/R ratio for the Whitehorse batholith may be associated with its emplacement history and location. Incorporation and assimilation of much colder country rocks as was discussed above, cools and crystallises the magma, more rapidly than would otherwise be the case, as the required heat transfer comes from the crystallising phases. Thus, if Morrison's basaltic assimilation scenario occurred,

the thermal energy required to drive a large convective hydrothermal system might not have been available. Instead, water circulation would be mostly restricted to highly permeable, structurally weak zones at the periphery of the intrusion and along crosscutting faults and fractures within the pluton.

The Mt-McIntyre Granophyre

The Mount McIntyre intrusion is an example that most closely follows H.P. Taylor's geological, petrographic and isotopic type-features of hydrothermally altered granitoid rocks. Its average whole-rock $\delta^{18}O$ of $+0.5^{\circ}/_{oo}$ indicates that the pluton is one of the most altered rocks in this study. An odd characteristic of the samples, though, is the normal quartz-magnetite and quartz-hornblende fractionations, that yield concordant isotopic temperatures of 640 to 680°C (Friedman & O'Neil, 1977). The temperatures are close to the solidus of natural and synthetic granodiorites at approximately 2 Kbar of H_2O pressure (Whitney, 1975, Piwinski & Wyllie, 1970). Normally, even in fully concordant systems (including feldspars), retrograde isotopic exchange in plutonic rocks does occur, yielding temperatures in the range 500 to 600°C, well below the solidus.

It is thought that the key to the apparent contradiction between petrographic observations and isotopic data lies in the separation and purification procedures used

to obtain the material for laboratory analysis. The procedures were made especially difficult by the small grain size of the constituent minerals.

Hornblende crystals in the granophyre (WHA 6a) are in the form of fine, euhedral needles that show chloritization and oxide formation at the periphery of the crystals. Extensive crushing and separation at the University of British Columbia isotope laboratory has produced a very clean hornblende fraction, suitable for K-Ar analysis (Morrison *et al.*, 1979) and in doing so eliminated the most brittle (and most ^{18}O exchanged) material at the edges of the minerals. The resulting $\delta^{18}O$ of hornblende is close to the +5 to +6°/₀₀ range usually encountered in granodioritic rocks (Bottinga & Javoy, 1975). Quartz does not recrystallise to secondary phases during ^{18}O exchange, therefore, normal separating techniques would not segregate the depleted edges of a quartz crystal from the core having a primary isotopic composition. The total quartz depletion in this sample, is estimated at 1 to 2°/₀₀, based on the four to one exchange ratio of figure 3-1. Thus the combination of purified hornblende and depleted quartz has, it is thought, produced the unrealistically high isotopic temperature.

Magnetite occurs as irregular blebs in the samples, closely associated with the replacement of hornblende. A similar explanation is likely for the high isotopic temperatures produced by the quartz-magnetite pair. Limited

crushing to relatively large mesh sizes (<100) was done initially to separate magnetite with the help of a magnet. The coarsest, most ^{18}O exchange-resistive material, was collected this way, leaving behind most of the fine-grained alteration-rim magnetite. The near normal $\delta^{18}O$ of magnetite (+1.9°/₀₀) and the concordant temperatures in sample WHA 6a (table 3-1) are thought to be fortuitous and do not reflect the final equilibration temperature of these mineral pairs. Separating biases were also encountered in sample WHA 6f and show some of the difficulties in obtaining high quality material for analysis from small samples of very fine grained (<1mm) altered rocks. The apparent equilibrium fractionation between quartz-feldspar, in a rock so obviously altered below the solidus, can only be due to impure separates caused by the ubiquitous myrmekitic and micrographic textures observed in the rock which rendered normal separating techniques useless. Abnormal oxygen yields for the separates confirm this view.

Despite certain difficulties in mineral separation, the relatively low whole-rock $\delta^{18}O$ values of this suite as well as a large $\Delta^{18}O$ for quartz-feldspar in WHA 6a clearly indicate that meteoric hydrothermal alteration has strongly affected this pluton.

Size effects on ^{18}O exchange

The effect of mineral grain size on reaction rates in isotopic and chemical exchange reactions has been noted in

experimental conditions (O'Neil & Taylor, 1967, Clayton et al, 1972) and in the field (Forester & Taylor, 1972, 1977). The increased reaction rates were ascribed to the fact that isotopic exchange occurs most readily at grain boundaries and therefore smaller grains are affected more rapidly because of their higher surface-to-volume ratios.

A plot of whole-rock $\delta^{18}O$ versus the average feldspar grain size is shown in figure 3-9 for the Whitehorse Trough samples. As is evident in the diagram, coarser rocks tend to be richer in ^{18}O. The interpretation of this trend can be done at two different levels, the first at the outcrop or batholith scale, the second, regionally.

The most convincing example of size-controlled rate of exchange is WHA 6a. Among the Mt-McIntyre specimens, WHA 6a is the most coarse-grained and exhibits relatively ^{18}O-rich quartz and magnetite. Zoned euhedral plagioclase (2.5 to 3 mm diameter) is rimmed with clay mineral alteration, while the core is intact (figure 3-10). Conversely, the plagioclase crystals in the finer grained samples of this suite, exhibit clouding, sericitisation and clay formation throughout the minerals (figure 3-11). Similarly, most of the magnetite in WHA 6a is an order of magnitude larger than in the other samples (>1mm versus <0.3mm) and is 3.4‰ richer in ^{18}O. The differences in whole-rock $\delta^{18}O$ values cannot be explained by modal abundance differences in the finer grained samples; for example, WHA 6e contains more quartz and less hornblende than WHA 6a which should give it

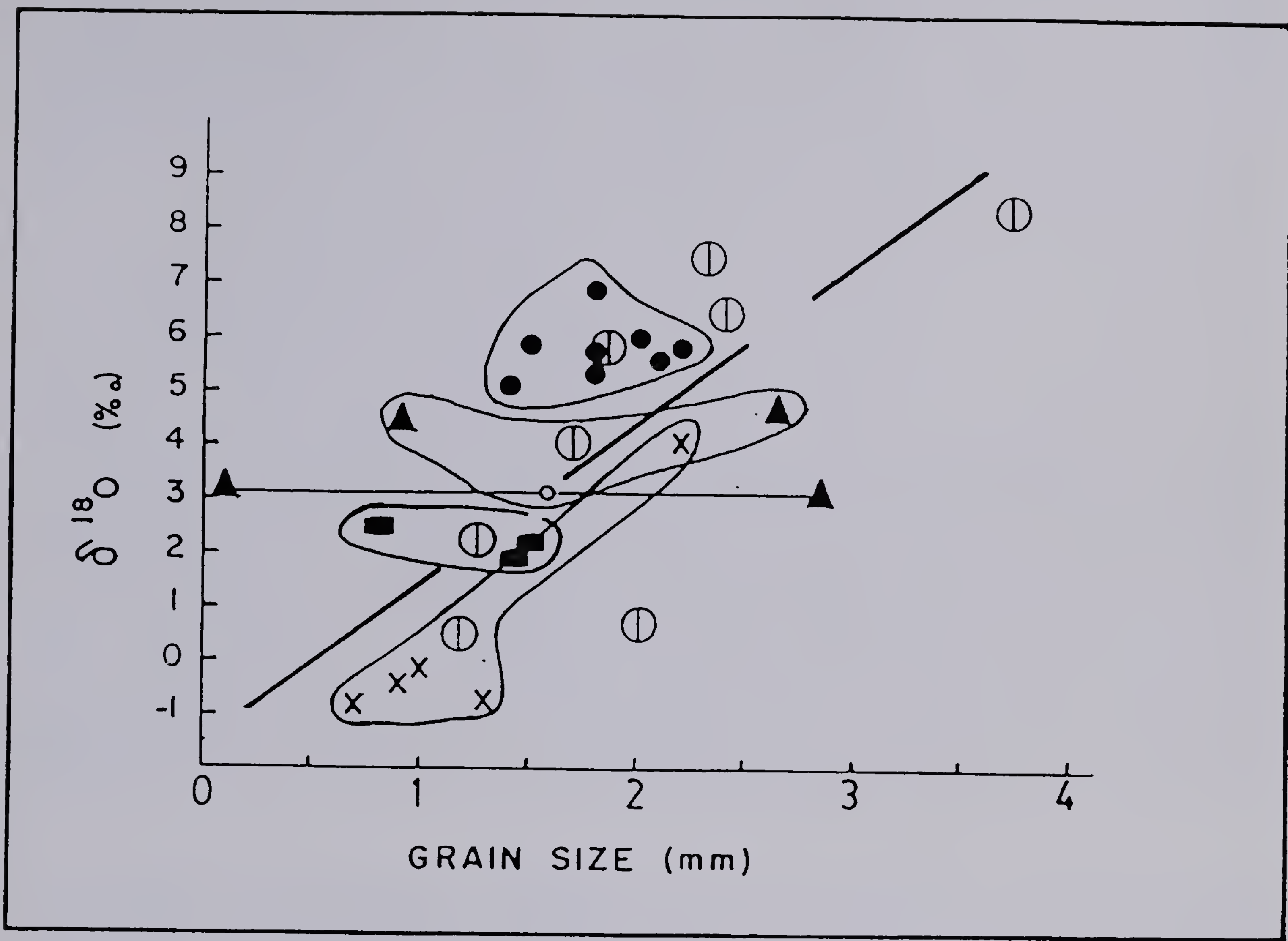
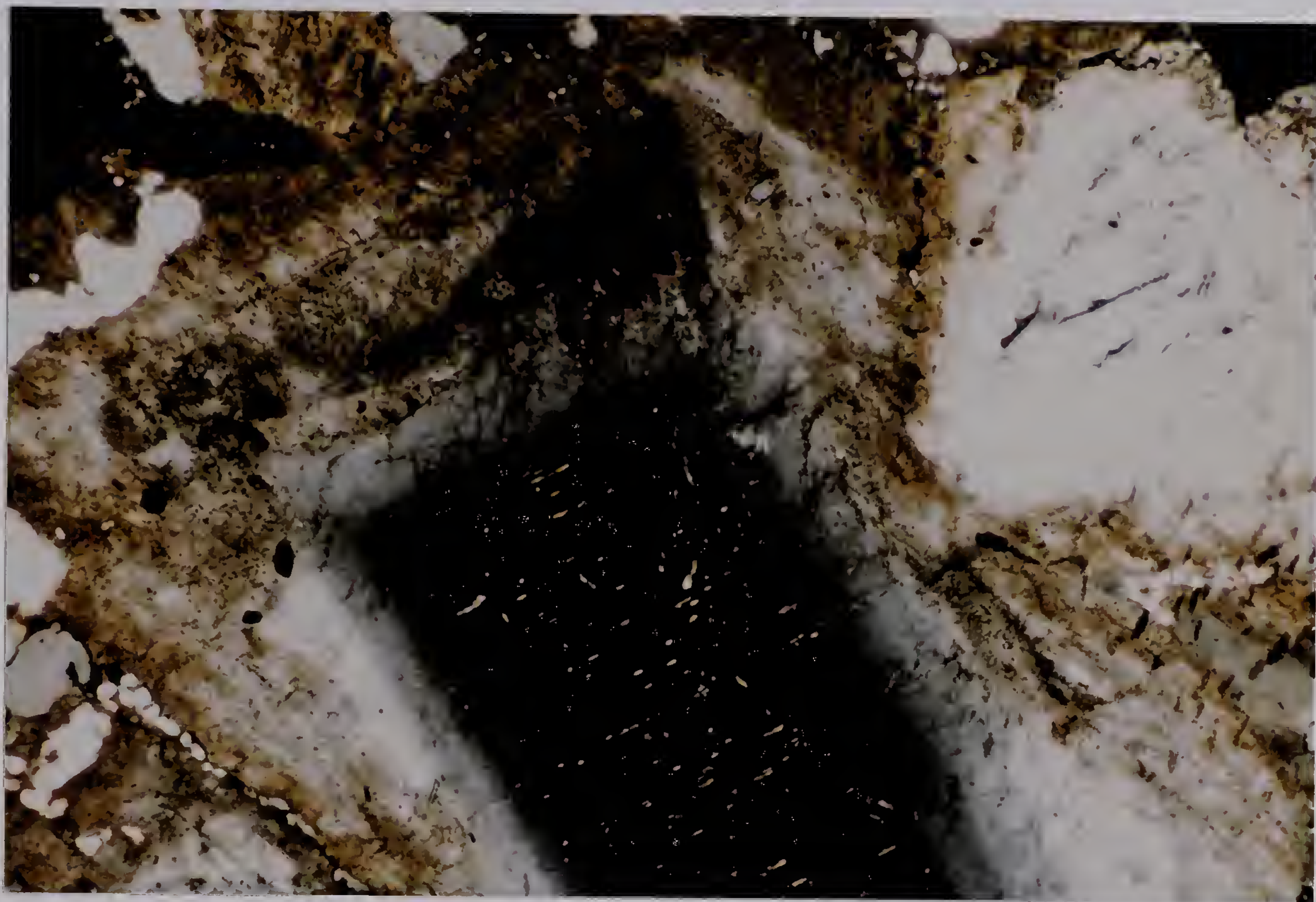
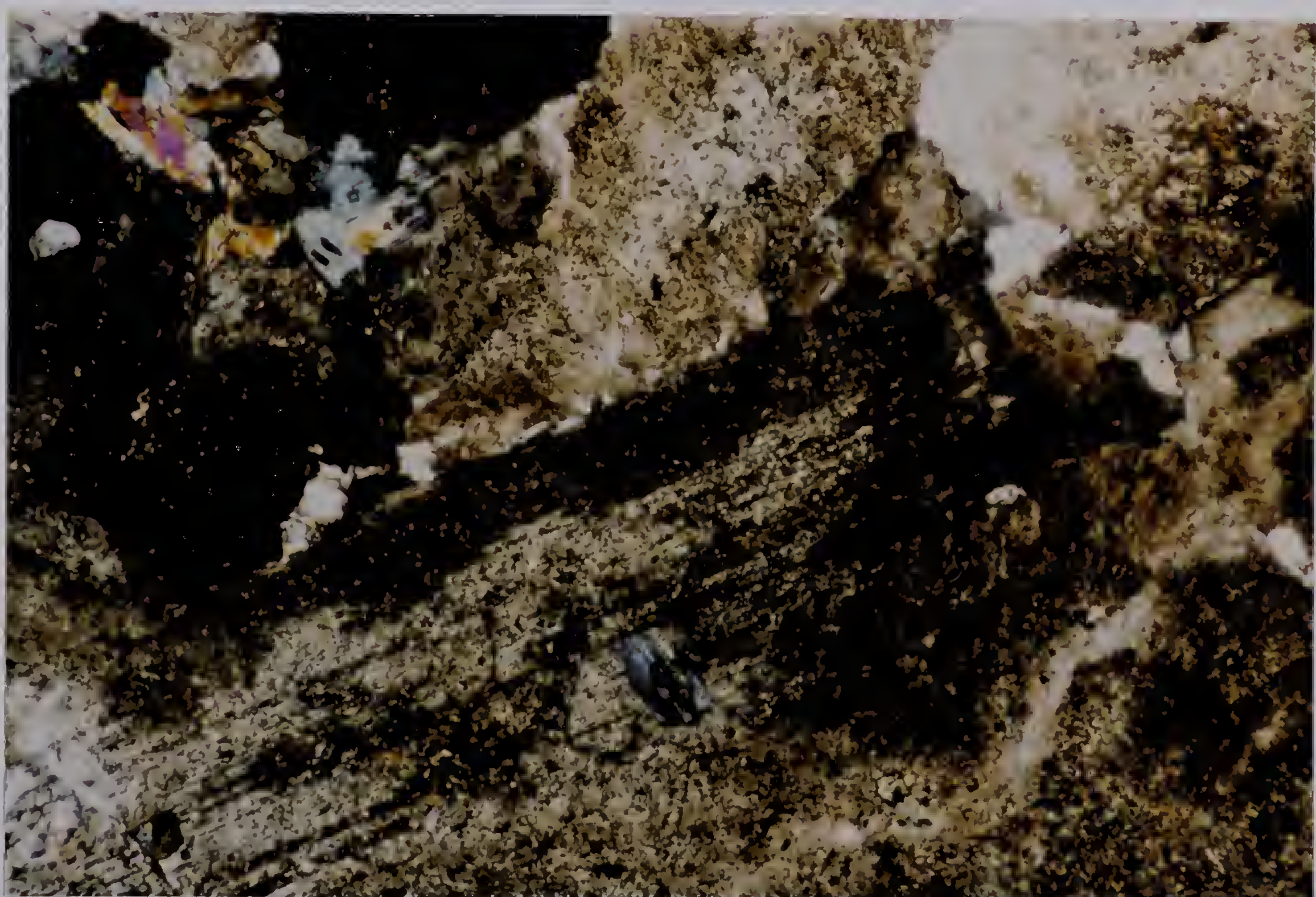


Figure 3-9 <u>Plot of whole-rock $\delta^{18}O$ versus feldspar grain size for the granitoid samples.</u> Large bisected circles are averages of suites or individual samples from discrete plutons (in the case of single sample from a plutonic body). Other symbols are individual samples from different suites; dots, WHB suite, triangles, WHA 3, rectangles, WHA 2, and Xs, WHA 6. The thick dashed line through the data points is a least-squares regression line with a correlation coefficient of 0.78 for the 8 bisected circles. The tie-line joining two triangles indicates that the sample is porphyritic and the small circle is the calculated average grain size of the feldspar in that sample.

Figure 3-10 Thin section of WHA 6a in crossed nicols (X 31). Central crystal is zoned plagioclase with altered margins. Alteration products in this sample are mostly clay minerals. Sericite flakes are seen within the central part of the plagioclase. Field of view spans 3.3 mm.

Figure 3-11 Sample WHA 6c in crossed nicols (X 31). Fully saussuritised plagioclase is in the center of the photograph. Chlorite and epidote alteration products are seen in the left-hand top corner. This sample is completely altered and of a finer grain size than WHA 6a.





a heavier oxygen content than the coarser sample. All samples were collected very close to one another at the center of the intrusion (figure 3-1 and appendix 2) and so changes in alteration intensity related to the position of samples (*i.e.* near contacts or faults) are not considered important for this suite.

On the regional scale, the positive trend in $\delta^{18}O$ versus grain size can be interpreted as a reflection of cooling rates and their relationship to the type of heat flow regime established during emplacement of the intrusive. Most of the plutons in Whitehorse area intruded upper crustal sedimentary rocks under broadly similar geological conditions. Sampling altitudes range from 2500 feet above sea level to 5000 feet. No correlation exists between altitude and $\delta^{18}O$. If Morrison's estimates of overlying sediment thickness (9000+ feet) are correct, then even by using exaggerated geothermal gradients such as 30°C/km (Beane & Titley, 1981), the temperature of surrounding country rocks at the time of intrusion would be below 100°C. The differences in country rock temperatures produced by altitude effects (<25°C) would be insignificant when compared to the 700+°C temperature gradient between magma and country rocks.

Convective cooling has been shown to be far more efficient in dissipating heat than simple conductive heat transfer (Cathles, 1977). Numerical modeling offluid flow through the Skaergaard intrusion (Norton & Taylor, 1979)

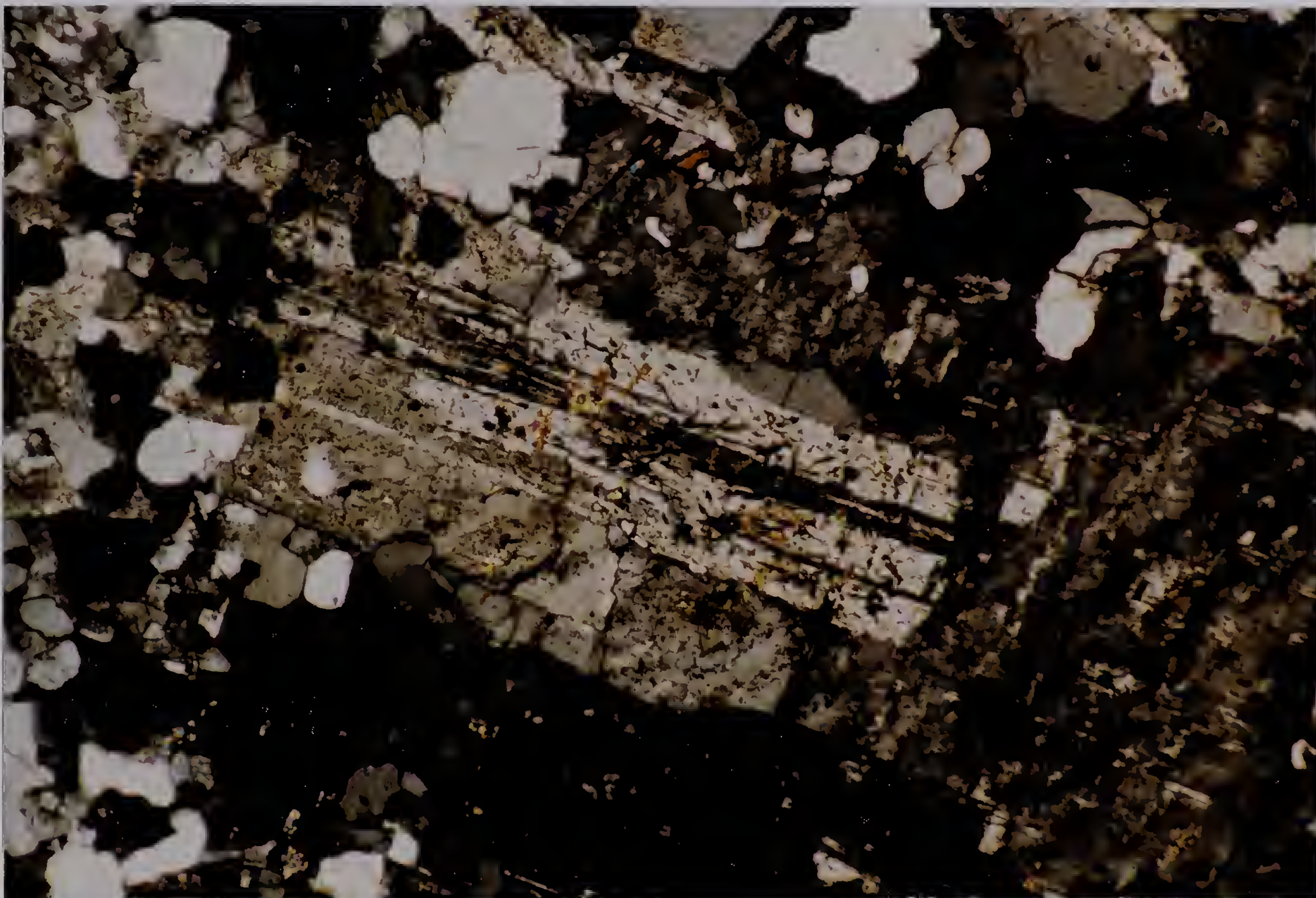
showed that a convective-conductive heat regime produced a cooling rate double that of conductive cooling alone. This difference in rate should affect crystal growth and the final grain size obtained for the pluton. Inasmuch as convective cooling is accomplished by fluid movement outside and within the intruding body, the association between rapid cooling, convective hydrothermal circulation and finer grain size should be expected. If the hydrothermal fluids were depleted in ^{18}O, as was shown above, the finer grain size rocks should also have lower concentrations of ^{18}O. The above observations break down in the rocks of the Coast Mountains and the Yukon Crystalline Terrane. This is explained by the fact that these regions are composed of granitoid rocks of very different settings (deep seated to epizonal), and therefore, cannot be compared in the same way as the Whitehorse epizonal plutons.

C. Comparison of the western terrane rocks with eastern terranes

On the basis of their mineralogical similarity with altered rocks of the Whitehorse Trough and the other western terranes (compare figures 3-12 and 3-13 with previous thin section photographs), the relatively high $\delta^{18}O$ values obtained for the granites of the Atlin Terrane and the Omineca Belt are anomalous. The observed results may be related to the ^{18}O composition of the fluids that circulated through the intrusives during the alteration phase.

Figure 3-12 Thin section of Kgal-1 in crossed polars (X 31). Moderate clay and sericite formation is seen on the central plagiocla-se crystal. White minerals are quartz, red brown → biotites. Field of view spans 3.3 mm.

Figure 3-13 Sample Mgdn-7 in crossed polars (X 31). Highly pronounced alteration is seen throughout the sample; central upper dark zone is serpentine after pyroxene, deep violet mineral is chlorite with inclusions of epidote (bright yellows). Lower part of the photograph is occupied by fully sericitised feld-spar.



Fluids in equilibrium with the K-feldspar ($\delta^{18}O$ = $+9°/_{oo}$) of the Surprise Lake Batholith sample, Kgal-3 (taken from a vein-filled fracture face), at temperatures of 200-400°C, would have an approximate composition of -0.6 to $+7.0°/_{oo}$ (O'Neil & Taylor, 1967). This is significantly richer in ^{18}O than the values calculated for vein fluids in the Whitehorse Batholith (-16 to $2°/_{oo}$). The relatively undisturbed whole-rock and mineral $\delta^{18}O$ observed in both the Atlin area and the Omineca Belt, could have resulted from equilibrium with the latter fluids. The difference in hydrothermal fluid composition cannot be the result of latitudinal and altitude effects since both the Whitehorse Trough and the Atlin Terrane are considered to have been adjacent to one another in Early Cretaceous time and probably even earlier (Monger, 1975, Bultman, 1979). More likely, the meteoric water was enriched in ^{18}O by interaction with marine metasedimentary rocks that crop out in the area.

Chert, limestone and argillite typically, have $\delta^{18}O$ ranging from +25 to $36°/_{oo}$ (Degens & Epstein, 1963, Knauth & Lowe, 1978), +18 to $28°/_{oo}$ (Degens & Epstein, 1964, Keith & Weber, 1964) and +14 to +26 (Savin & Epstein, 1970a, Savin, 1980, Longstaffe *et al.*, 1982). These $\delta^{18}O$ values are significantly higher than the $\delta^{18}O$ of immature volcanic and volcaniclastic rocks (+8 to $+15°/_{oo}$) observed in this study and those reported in other papers (Savin & Epstein, 1970b, Magaritz & Taylor, 1976c, Longstaffe & Schwarcz, 1977). At

average temperatures of 50 to 150°C, compatible with estimated burial depths (Morrison, 1981), water in equilibrium with the above marine metasediments would have compositions ranging from +5 to 16°/₀₀, +1 to 11°/₀₀ and 0 to +12°/₀₀ respectively (Matsuhisa *et al.*, 1979, O'Neil & Epstein, 1966b, Savin, 1980). The ranges are in fairly good agreement with the estimated $\delta^{18}O$ of fluids in equilibrium with sample Kgal-3.

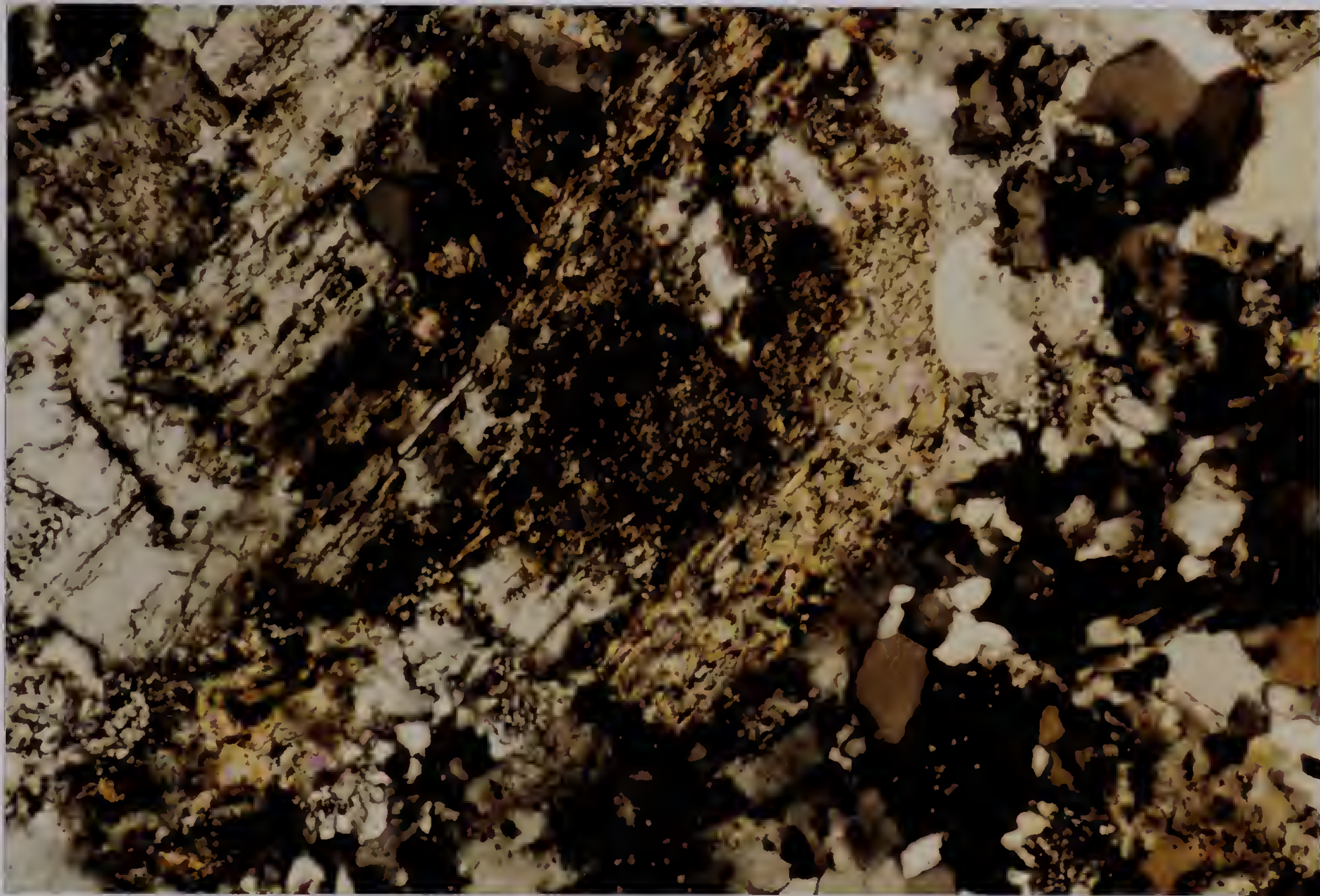
Locally, as in samples Mgdn-7 and JKqm-2, a lowering of $\delta^{18}O$ of about 2°/₀₀ has been observed along with thorough propylitisation and saussuritisation (figure 3-13). The relatively high $\delta^{18}O$ content of these rocks (+9.2°/₀₀) despite hydrothermal alteration tends to confirm the high $\delta^{18}O$ groundwater hypothesis. Although specific comment was made about the Omineca Belt granites, similarities in alteration products (figure 3-14 and 3-15) and country rock types of both terranes and $\delta^{18}O$ data of table 2-5, show that a similar process could have affected these rocks.

D. Potassium-Argon ages and ^{18}O concentrations

Biotite K-Ar ages are known to be susceptible to resetting by metamorphic heating and are affected by variable cooling rates in solidifying plutons. Until recently, the role played by alteration on K-Ar age determinations has been somewhat neglected in the scientific literature. The use of oxygen isotopes in investigating the possible effects of hydrothermal heating and alteration

Figure 3-14 Thin section of sample 1075 in crossed polars (x 31). Strong sericitisation can be seen on the central plagioclase crystal. Epidote in trace amounts (lower right hand corner) and muscovite (high order pinks and yellows) are also present. Field of view is 3.3 mm.

Figure 3-15 Thin section of sample KTP in crossed polars (X 40). Moderately altered sample. Large plagioclase crystal is strongly recrystallised to sericite. Brownish clay mineral shading can be observed throughout the sample. Muscovite (high order colors) and quartz (white and brown [almost at extinction]) are fresh. Field of view is 2.6 mm.



could prove to be a worthy tool.

Weathering of biotites of a Precambrian crystalline basement in the Chadian Republic of Africa (Clauer, 1980, Clauer et al, 1982) systematically decreases the apparent ages of the Rb-Sr and K-Ar geological clocks. The authors indicate that both ^{87}Sr and Ar are leached out during the weathering process and that the two radiometric systems mimic each other, yielding similar progressively younger ages throughout the alteration process.

In the Idaho Batholith, biotite K-Ar ages have been reset from Mesozoic time (90 Ma) to the Tertiary Period (50 Ma) by leucocratic epizonal granitic intrusions (Armstrong, 1974, 1975, Criss *et al.*, 1982). The relative effects of regional uplift, hydrothermal heating and mineralogical alteration were evaluated by Criss and others (above). Their study showed that close to the intrusive centers the largest effect was caused by hydrothermal heating and alteration while further away from the plutons, effects of progressive uplift and doming predominated. The younger ages of lower altitude samples indicated to the authors above, that some of the observed age patterns were caused by the samples passing through the biotite K-Ar closure temperature at different times.

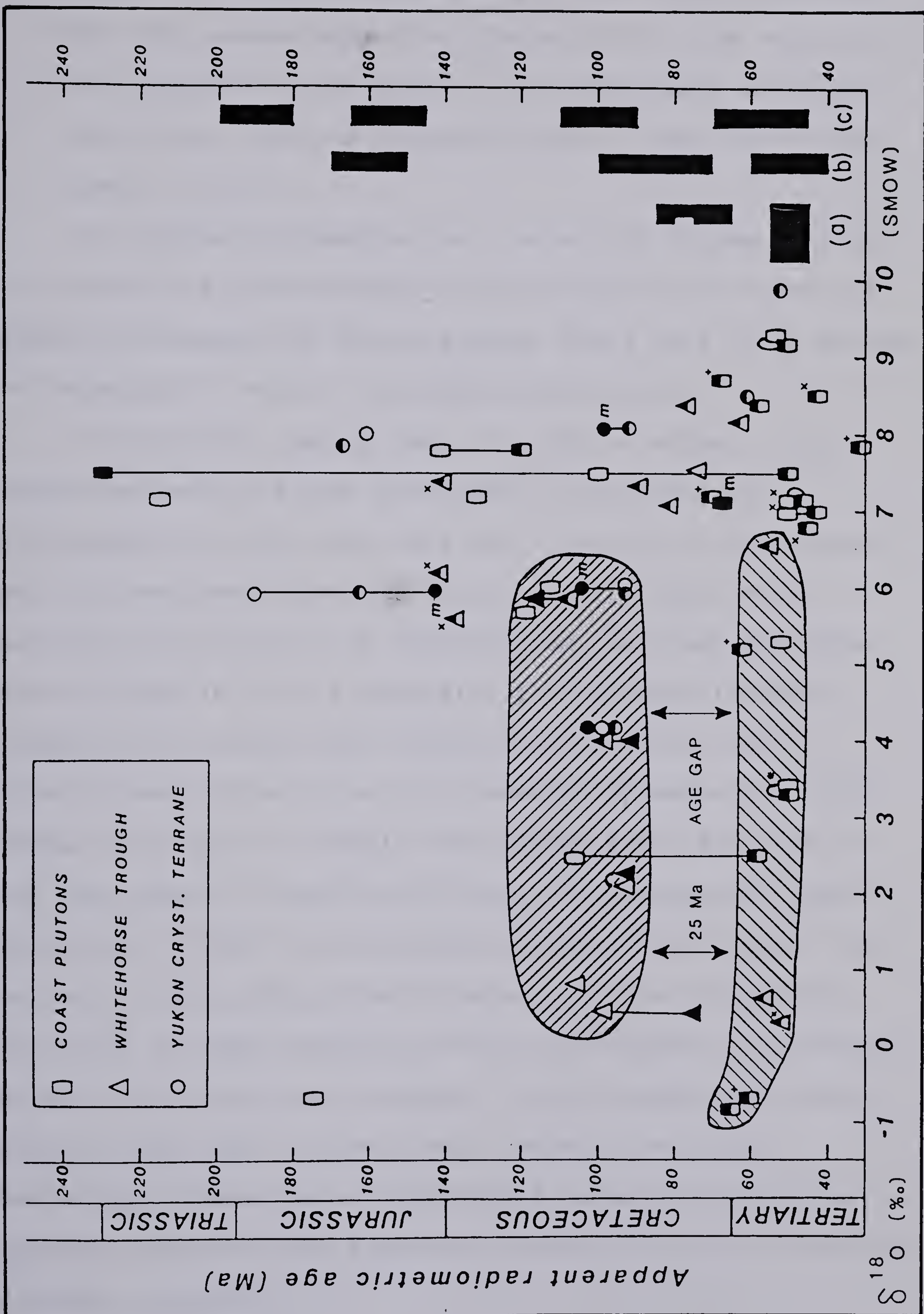
Oxygen isotope results of whole-rock samples from this study can not be directly compared to K-Ar ages in the same way as the two previous studies since our approach is regional, not local in scale. Comparison of the two isotopic

systems does, however, yield interesting results. Figure 3-16 utilises whole-rock $\delta^{18}O$ values from this study and those from the Skagway and Prince Rupert - Prince George traverses (Magaritz & Taylor, 1976a, 1976b) with K-Ar ages from published (Wanless & Lowdon, 1963, Lowdon, 1961, Tempelman-Kluit & Wanless, 1975, Hutchison, 1970, White *et al.*, 1970, Harrison *et al.*, 1979, Morrison *et al.*, 1979 and Bultman, 1979) and unpublished sources (R. Armstrong).

Surprisingly narrow age bands are observed in association with hydrothermally altered rocks, as is shown in the diagram. The overall features of the plot can be summerized as follows;

1. Fairly continuous plutonism has occurred in the Whitehorse Trough and the rest of the Intermontane Belt since the Late Jurassic epoch to Eocene time.
2. More episodic plutonism seems to have occurred in the Coast Plutonic Belt and the Yukon Crystalline Terrane throughout that same period although this might be due to limited sampling in these areas.
3. About half of the sampled plutons in the Whitehorse area have ages in the 115 to 90 Ma period. Most of these rocks were altered by meteoric water. Dated rocks from the Prince George area, B.C., are also isotopically altered but their ages are considerably older (≅ 140 Ma) than the granites of this study (White *et al.*, 1970, Magaritz & Taylor, 1976b).
4. Most of the CPC rocks have biotite ages in the Eocene

Figure 3-16 Apparent radiometric age versus whole-rock $\delta^{18}O$ of granitic samples from the three western terranes. Filled symbols, Rb Sr whole-rock isochron, filled symbol with an *m*, mineral-rock isochron. Half-filled symbols, biotite K-Ar age, open symbols, hornblende K-Ar age. "x" near a symbol indicates the the sample is from the Prince Rupert - Prince George traverse (Magaritz & Taylor, 1976b). A "+" shows the sample comes from the Skagway traverse (Magaritz & Taylor, 1976a). The shaded areas incorporated altered samples. The vertical bars on the right hand side of the diagram are;(a) a schematic histogram of K-Ar ages from porphyry copper deposits in central B. C., (b) periods of major plutonic activity as described by Pëto based on K-Ar ages (1974), and (c) periods of climactic plutonism as stated by Roddick and others (1975). The tie-lines join samples on which multiple dating methods were applied.


COAST PLUTONS
WHITEHORSE TROUGH
YUKON CRYST. TERRANE
AGE GAP
25 Ma
(a) (b) (c)
(SMOW)
TERTIARY
CRETACEOUS
JURASSIC
TRIASSIC
Apparent radiometric age (Ma)
δ18O (‰)

period (50-60Ma). These dates are considerably younger than the assumed ages (Late Paleozoic to Late Mesozoic) for the majority of rocks in the Coast Range batholith (Hutchison, 1970, Gabrielse & Reesor, 1974, Armstrong & Runkle, 1979).

An interesting observation, concerning figure 3-16 is the absence of hydrothermally altered rocks in between the Middle Cretaceous and Eocene groups. There is a 25 Ma period or "age-gap" in which no altered samples plot.

Interpreting the age data in light of oxygen isotope ratios necessitates some knowledge of the geological environment in which the rocks are situated. In the western part of the Coast mountains, Prince Rupert area, many plutons are thought to be derived from the crust at depths greater than 15 to 20 kilometers. Their oxygen isotopic composition reflects equilibrium between constituent minerals and little or no $\delta^{18}O$ lowering by meteoric fluids (Magaritz & Taylor, 1976b). Their young ages (43 to 50 Ma) are the result of uplift and erosion of the mountain chain (Hutchison, 1970). At the quoted depths of emplacement and anatexis, in a crystalline terrane, it is not surprising that most of these samples have not interacted with meteoric water in any major way. However, in the central and eastern parts of the Coast Plutonic Belt, meteoric water has definitely interacted with granitoid rocks during Tertiary epizonal plutonism and explosive volcanic activity (Magaritz & Taylor, 1976a,b).

Despite the pervasive alteration present in the area, the Whitehorse Trough biotite and hornblende K-Ar and whole-rock Rb-Sr radiometric ages are regionally consistent. However, no biotite-hornblende K-Ar pairs are available for these rocks and comparison of whole-rock Rb-Sr with mineral K-Ar dates (bio or Hbl) yields inconclusive results. In most cases, because of the lack of major compositional variation in the rock suites and the presence of isotopic heterogeneities in some samples (*i.e.* WHA 6 a-f), the small number of samples used in the Rb-Sr isochron regressions produces large errors in age data (Morrison *et al.*, 1979). The magnitude of the errors seems to be linked to the amount of $\delta^{18}O$ lowering of the samples (Dagenais, 1983). The most altered samples show the most discordant radiometric dates (K-Ar versus Rb-Sr) and are generally associated with younger ages. Because of these problems, it is difficult to determine the exact nature of the Middle to Late Cretaceous radiometric dates of the plutonic suites of the area.

Gabrielse and Reesor (1974) have suggested that most of the plutons cropping out on the eastern margin of the Coast Plutonic Complex and the western side of the Intermontane Belt were of Early to Mid-Jurassic age based on results in southern British Columbia (Nguyen *et al.*, 1968, Northcote, 1969, among others). The younger ages found in the Whitehorse Trough and the presence of the "age gap" in figure 3-16 could be the result of resetting by hydrothermal heating and alteration at temperatures above 300°C. Uplift

effects on K-Ar dates are not important in this area since realistic estimates of crustal depths <3 km and pressures (1 Kbar) indicate that the enclosing rocks should not have been heated by burial above the closure temperature of Ar in biotite (about 200-300°C).

Convective heat dissipation through fluid movement has been shown to be a very efficient way of cooling a plutonic mass after its emplacement. Hydrothermal circulation involves larger volumes of rock (Taylor, 1974a,b, 1977, etc.) heated at low to moderate temperatures (200 to 400°C) for longer periods of time than would be the case for simple conductive cooling alone (Cathles, 1977). As was discussed in the previous chapter, cooling rates were modelled to be twice as rapid in convective-conductive regimes than in conductive-only situations (Norton & Taylor, 1979). Reynolds (1980) estimated that at a distance of up to 1 km from the margin of 2 km wide circular stocks in the Sierrita porphyry copper deposit (Arizona), the temperatures of the older host granitoid rocks and associated fluids were approximately 300°C. The hydrothermal aureole around the Red Hills granite of Skye, Scotland was at least as large as the Sierrita deposit and hydrothermal temperatures may have been present up to 10 km from the periphery of the intrusion (Forester & Taylor, 1977).

On the basis of the above facts, it would appear logical that if enough plutons invaded a certain volume of permeable rock in the upper crust in a short time interval,

then the whole volume would be reset to the age of the plutonic emplacement through hydrothermal heating. The position, in time, of the age groups associated with low-^{18}O rocks tends to confirm the above hypothesis, since they correspond to the periods of climactic plutonic activity as seen through the abundance of K-Ar biotite and hornblende ages along the entire length of the Canadian Cordillera (Gabrielse & Reesor, 1974, Roddick *et al.*, 1975).

Two Coast Range samples from this study, one unaltered, LW77 A58B, from the Mt-Caplice granodiorite and one altered to a $\delta^{18}O$ value of +2.5°/₀₀ (RW-371), independantly confirm the occurrence of thermal resetting in the Coast Mountains. In both cases, resetting of biotite K-Ar age to 50 Ma has occurred while both samples contain hornblende yielding 100 Ma dates. A Rb-Sr four point isochron for the Mt-Caplice pluton yields an age of 230 Ma (R. Armstrong, unpublished data) while a single point whole-rock Rb-Sr date for the foliated granodiorite-granite of the Ram Property gives a maximum age of 200 Ma, assuming a low initial $^{87}Sr/^{86}Sr$ ratio of 0.705 (Watson *et al.*, 1981). Therefore, both these samples are probably of Early to Middle Mesozoic age and have had their K-Ar clocks reset to the Middle Cretaceous and Eocene plutonic events.

In the Whitehorse Trough, such a conclusion can only be supported by comparison with other intrusions further south in the Intermontane Belt that show concordant Rb-Sr and K-Ar of Middle Jurassic age. This is true of the normal $\delta^{18}O$

Guichon Creek Batholith and other granitoid plutons of the area (Gabrielse & Reesor, 1974 , Taylor & Magaritz, 1978, Preto *et al.*, 1979). The resetting hypothesis for the altered batholiths of the Whitehorse is speculative since no biotite-hornblende pairs are available and because of the general agreement of whole-rock Rb-Sr "errorchrons" with the K-Ar data. Isotopic heterogeneities of radiogenic isotope systems have been observed previously in the hydrothermally-altered plutonic rocks Armstrong *et al.*, 1977, Criss & Taylor, 1982, Forester & Taylor, 1977) but at present this association has only been empirically linked. In view of the ambiguities above and the lack of independent age data, the exact nature of the 110-90 Ma age group in the Whitehorse Trough is still to be determined.

E. Summary

From the discussion above, it is clear that, in the study area in general, and more specifically in the Whitehorse Trough, meteoric-hydrothermal alteration was pervasive in character. Except for the WHB suite which may have intruded as a low ^{18}O magma (+6.5°/₀₀?), all other granitic samples with $\delta^{18}O$ values of less than ≅ +6.5°/₀₀, have suffered alteration by ^{18}O-depleted groundwaters. In some mafic end-member rocks, the ^{18}O depletion is not associated with secondary hydrous mineral formation because of higher than average, >400°C temperatures of hydrothermal exchange. In the vast majority of cases though; sericite,

chlorite, clay minerals, epidote, calcite and secondary oxides are present after the alteration event. The above assemblages suggest a propylitisation temperature of less than 350°C. The estimated fluid ^{18}O composition ($-16^{o}/_{oo}$) is comparable to other studies of hydrothermal fluids in the Cordillera. Water-rock ratios in the Whitehorse area are high (up to 0.9) and similar to values found in other epizonal plutons in volcanic terranes. In the other areas of this study, comparable amounts of water must have cycled through the rocks, where physical conditions permitted it. In some cases, as in Tqm-1 from the Teslin Suture Zone ($\delta^{18}O = -2.6^{o}/_{oo}$) and in the Whitehorse skarns ($\delta^{18}O \leq -3.4^{o}/_{oo}$), the water-rock ratio must have exceeded 1.

The $\delta^{18}O$ differences observed between mineralogically-altered rocks of the western and the eastern terranes is mostly due to the isotopic composition of fluids circulating in the respective regions. The relatively high $\delta^{18}O$ groundwaters in the latter terranes may have acquired their isotopic signature by low temperature (50-150°C) interaction with the high $\delta^{18}O$ country rocks of the area prior to their infiltration into the batholiths.

The association between ^{18}O depleted rocks and narrow, Middle Cretaceous and Tertiary age bands, suggests that hydrothermal heating has reset the K-Ar clocks of plutonic rocks throughout the Whitehorse Trough and in the Coast Plutonic Belt.

IV. ISOTOPIC FEATURES OF UNALTERED ROCKS

Up to the present, hydrothermal alteration has been the main focus of discussion. This section will deal with primary $\delta^{18}O$ features of unaltered granitoid rocks and the surounding country rocks. will be discussed. The large variations in isotopic composition between rocks of different terranes will be related to their petrogenetic history and to magmatic source regions. By comparing $\delta^{18}O$ with Sr isotope ratios, furthers constraints will put on possible genetic models.

A. $\delta^{18}O$ of granitic rocks

Firstly, one has to define the term primary. As was stated previously, the assumed mantle value for oxygen is equal to +5.5 to 6°/₀₀ (figure 4-1). Whole-rock $\delta^{18}O$ values below this limit were rejected because the samples featured disequilibrium mineral ^{18}O fractionations and mineralogical alteration. Values higher than 6°/₀₀ were not rejected unless their $\Delta^{18}O$ (qtz-fp) was larger than about +2.5; the largest equilibrium fractionation for calcium-rich plagioclase in granites. On this basis, fifty samples were accepted as reflecting primary isotopic characteristics, out of a total of 86 whole-rock analyses.

The results, when plotted by region (figure 4-1), show two distinct groups of rocks. Samples from the Whitehorse Trough and westward have similar ^{18}O ranges and averages, generally increasing from quartz diorites to true granites.

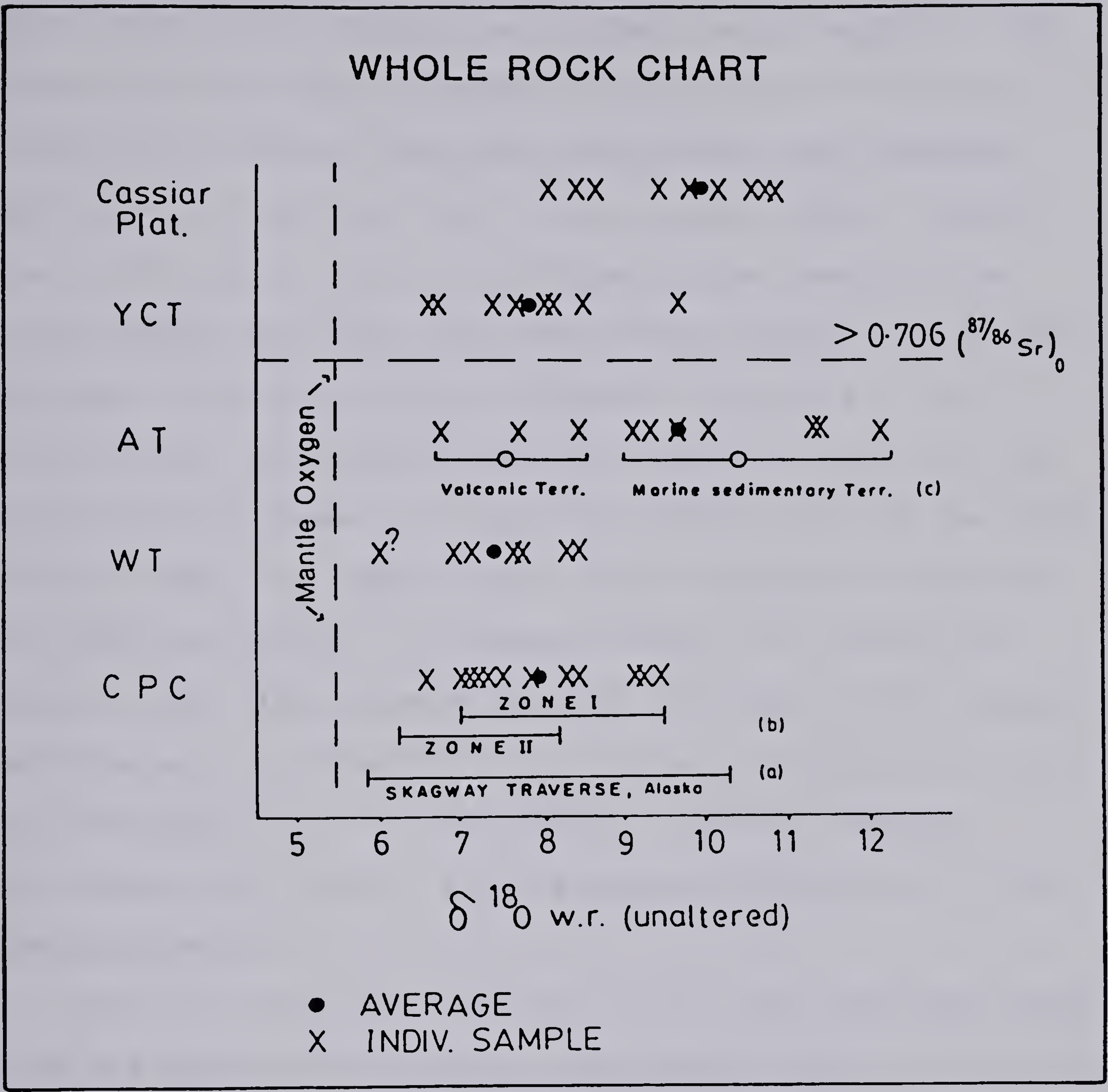


Figure 4-1 Primary $\delta^{18}O$ values for granitoid rocks of the Yukon. The results are subdivided into the five tectonic units that are referred to in the text. X - individual analysis; dot - mean value for the terrane; open circle, average isotopic composition of groups of rocks within the Atlin terrane. The 0.706 line is drawn to indicate that the two top terrane have high initial Sr ratios. (a) range in $\delta^{18}O$ found in granites of the Skagway traverse (Magaritz & Taylor, 1976a), (b) ranges corresponding to the respective zones of the Prince Rupert - Prince George traverse (Magaritz & Taylor, 1976b), and (c) country rock types surrounding the granites having a mean isotopic composition defined by the open circles.

This relationship breaks down in the western margin of the Coast Mountains where granodiorites have $^{18}O/^{16}O$ ratios identical or larger than quartz monzonites (see samples Mgd-4, -1, R3 in Table 2-3). In one sample, Mgd-3, quartz has a $\delta^{18}O$ of +11.3°/oo, the highest value found in the three western terranes. The Coast Range results are in good agreement with data from the Skagway (figures 4-1) and Prince Rupert traverses (Magaritz & Taylor, 1976a, b). The progressive increase in ^{18}O of the western part of the Coast Plutonic Belt was noted in the former studies, as shown by the shift to higher $\delta^{18}O$ values in Zone I of the Prince Rupert rocks. The presence of high ^{18}O rocks in the western part Skagway traverse (+9.7 to 11.5°/oo) indicates that the enriched Mgd-3 quartz (this study) is not an isolated occurrence, but rather is quite common in that part of the Coast Mountains.

The $\delta^{18}O$ averages of 7.4 to 7.9°/oo for granitoid rocks from the western terranes are relatively close to the mantle oxygen value. They would be classified as "I" types on the basis of mineralogical and isotope data (Chappell & White, 1974, O'Neil &Chappell, 1977). Calculated confidence intervals at the 0.1 level indicate that statistically, the rocks from the three western terranes have identical $\delta^{18}O$ averages. The significance of the $\delta^{18}O$ results can best be appreciated by comparing them to data from plutonic and volcanic arc-derived rocks from other areas of the world.

In the Izu-Mariana Arc of Japan, the Quaternary Tholeiitic (basaltic to dacitic) Hachijo-Jima suite varies in $\delta^{18}O$ from +5.7 to +6.7‰ while concurrently, silica content rises from 47 to 73% (Matsuhisa, 1979). $\delta^{18}O$ and silica analyses (Chivas *et al.*, 1982) of a Quaternary calc-alkaline suite of plutonic rocks from the island of Guadalcanal (South Pacific) also show similar ranges in composition (+5.7 to +7.2‰ and 49 to 75%, resp.). Both these suites are part of immature arcs built directly on relatively thin oceanic crust. It is thought that they are pristine mantle-derived rocks which could not have been contaminated by continental crust. Simple crystal fractionation is enough to explain the $\delta^{18}O$, bulk chemistry, trace element and Rb-Sr data (ref. cited).

Two recent studies of calc-alkaline intrusive rocks in continental settings, one in the Central Andes (Longstaffe *et al.*, 1983) and one in the Himalayas (Blattner *et al.*,1983), show $\delta^{18}O$ values ranging from +6.0 to +8.3‰ and from +6.0 to +7.6‰ respectively. These values are up to 1‰ richer in ^{18}O compared to the results found in the oceanic suites above, but are virtually identical to the $\delta^{18}O$ spread (+6.0 to +8.4‰) observed in the three western terranes of this study, if highly silicic post-tectonic granites and country rock-echanged granodioritic samples are not considered.

The older granitic rocks of this study are thought to be remnants of a Late Paleozoic-Early Mesozoic island arc of

similar tectonic setting as the present-day South Pacific oceanic arcs (Gabrielse & Reesor, 1974, Tempelman-Kluit, 1979). Younger intrusions are probably related to the collision of the aforementioned Stikinia arc with the ancient western margin of the North American plate and also to the reestablishment of Pacific plate subduction beneath the newly-accreted terranes (Godwin, 1975, Tempelman-kluit, 1979, Monger *et al.*, 1982). As such, taken as a whole, the granitoid rocks of the Yukon Crystalline Terrane, the Whitehorse Trough and the Coast plutonic Complex are a mixture of rocks of oceanic and continental character.

Their $\delta^{18}O$ values can be explained by progressive ^{18}O enrichment during fractional crystallisation similar to the Hachijo-Jima and Guadalcanal suites. The 2.4‰ enrichment from typical mantle values of 5.7 to 6.0‰ found in the more siliceous rocks of the study is larger than the observed $\delta^{18}O$ increase in the oceanic suites (1-1.5‰) possibly because of the lower temperatures and slower cooling rates in more deep seated continental plutonic rocks. The slow cooling rate should have permitted isotopic retogression to occur, which would have further increased the final $\delta^{18}O$ of the rocks beyond the normal effects of crystal fractionation, as was discussed by Longstaffe and others (1983) for the Central Andean rocks.

Alternatively, a small amount (<10%) of crustal contamination by continental material of +14 to +16‰ $\delta^{18}O$ composition could also produce a 1‰ increase in the final

$\delta^{18}O$ of magmas, above a strictly differentiation-induced 1 to 1.5‰ enrichment of the Hachijo-Jima type. However, as will be seen in a later section, this hypothesis is not favored by the Sr isotope systematics of the Coast Plutonic and Whitehorse Trough granitic rocks. Except for the silicic and country rock-exchanged samples noted above, the western terrane intrusions could not have substantially interacted with upper crustal, continent-derived high ^{18}O material.

Rocks from the two geological provinces east of the Whitehorse area, also have similar $\delta^{18}O$ values, which are significantly enriched in ^{18}O relative to those obeserved to the west. Note that even though some of the former rocks have been hydrothermally altered (chapter 3), the quartz values in table 2-5 indicate that their original high $\delta^{18}O$ character has been preserved during propylitisation. As such, the consideration of their $\delta^{18}O$ values in a chapter concerned with primary ^{18}O signatures is not considered to be a contradiction in terms.

A very large spread in whole-rock results (6.7 to 12.2‰) is seen in the Atlin Terrane rocks. The range spans both I and S types based on the classification of O'Neil and Chappell (1977) and O'Neil and others (1977). The final $\delta^{18}O$ of the granitoids rocks in the area appears to be controlled by the country rocks surrounding the intrusions. At the margins of the terrane, the three plutons that intrude mostly into volcanic and ultrabasic rocks of oceanic character have $\delta^{18}O$ values that range from +6.7 to +8.5‰.

Their size is an order of magnitude smaller than the large batholiths cropping out in the central part of the Atlin area. The latter rocks are enriched in ^{18}O (mean $\delta^{18}O$ >+10°/₀₀) and emplaced through a (thickened?) crust composed of marine sedimentary rocks. The Omineca Belt and Selwyn Basin granites are emplaced in Precambrian and Paleozoic shelf metasedimentary rocks and are also isotopically enriched ($\delta^{18}O$ of +8.0 to +12.7°/₀₀).

The $\delta^{18}O$ compositions of the eastern terrane rocks are comparable to the peraluminous High Himalayan leucogranites (+9.1 to +12.2°/₀₀) of Blattner and others (1983), the Hercynian Cornubian batholith (+10.8 to +13.2°/₀₀) of Sheppard (1977) and to the South Mountain batholith in Nova Scotia (+9.6 to +12°/₀₀ *in* Longstaffe *et al.*, 1980). Melting and assimilation of high-^{18}O metasedimentary rocks or pluton/contry rock isotopic exchange have been suggested by the above authors to explain the high $\delta^{18}O$ values found in the peraluminous granites mentioned. The isotopic characteristics of the eastern terrane rocks support a country rock exchange hypothesis, both for the high-^{18}O granites throughout the entire area and the small, relatively low-^{18}O stocks at the periphery of the Atlin Terrane. Partial melting of the surrounding country rocks is not a realistic explanation because of the low to intermediate metamorphic grades of the terranes, and the clearly intrusive nature of the plutonic contacts. Since melting of pelitic material can proceed at depths of the

order of 15 to 25 km (Wyllie & Tuttle, 1961, Green, 1976) the large intrusives of the Atlin , Omineca and Selwyn Basin terranes could be remobilised crustal melts from deeper levels of the continental crust. As will be discussed later, Sr isotopes tend to confirm the proposed continental origin of the eastern terrane granites.

B. Country rocks

Sampling of country rocks was done for two distinct reasons; to observe hydrothermal alteration effects surrounding plutons and to compare granitic $^{18}O/^{16}O$ ratios with those of rocks in the immediate area. Only in the Omineca, the Whitehorse and the Coast Plutonic belts is sampling sufficient to warrant comment (figure 4-2).

In the Coast Mountains, three schists sampled from the western margin of the belt (Table 3-2, previous chapter) have $\delta^{18}O$ values (10-13°/₀₀) that lie within the published ranges of highly metamorphosed pelitic rocks (Garlick & Epstein, 1967, Shieh & Taylor, 1969a, Longstaffe & Schwarcz, 1977). Samples analysed by Magaritz and Taylor (1976b) from the Central Gneiss Complex near Prince Rupert have lower ^{18}O values (6.8 to 9.5) compatible with Archean migmatites (Shieh & Schwartz, 1974, Longstaffe & Schwarcz, 1977, Longstaffe, 1979, Longstaffe & Birk, 1981, Longstaffe & Gower, 1983) and some granulitic rocks in old cratons (Wilson & Green, 1971, Fourcade & Javoy, 1973). Granulites are thought to be precursor to some of the anatectic

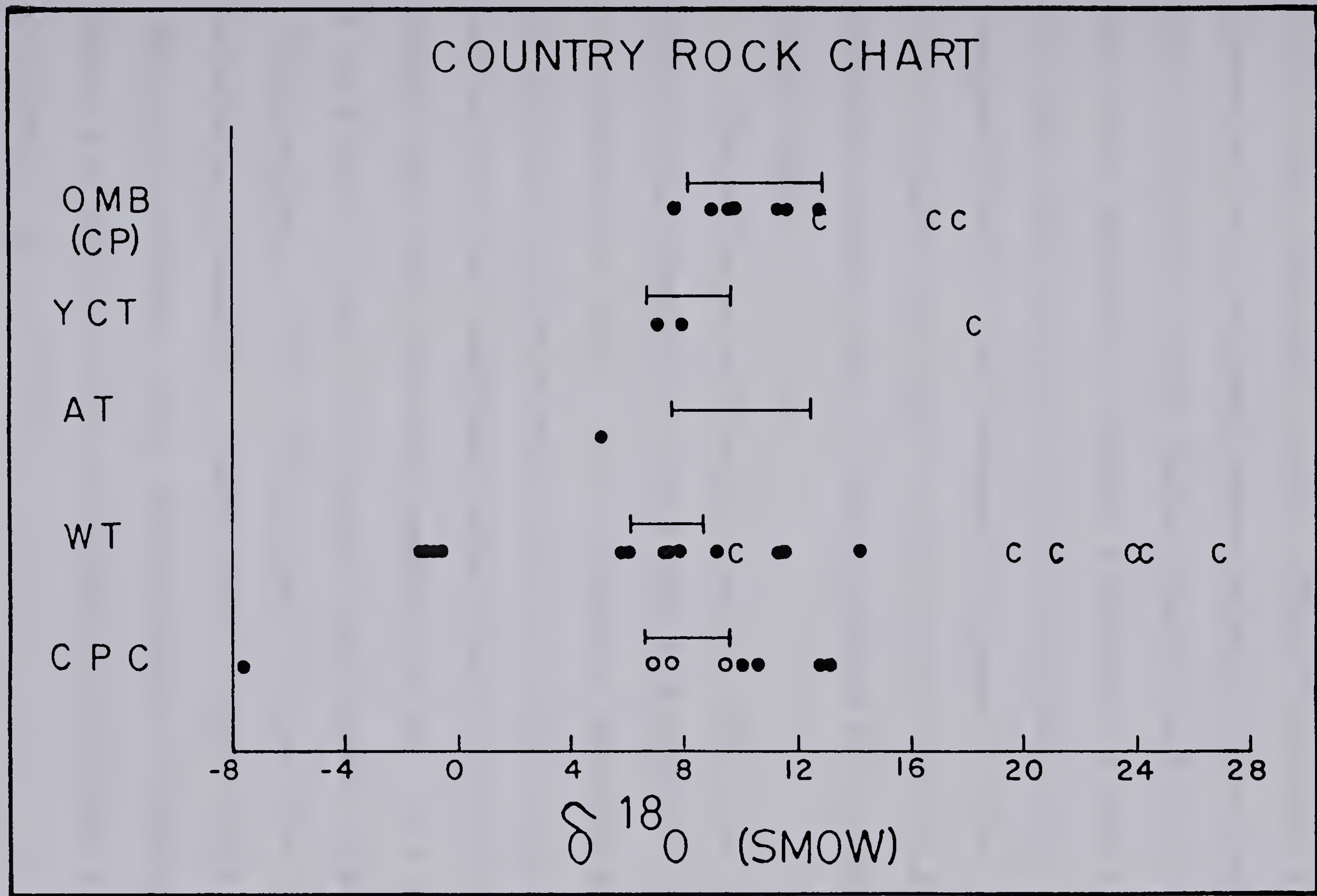


Figure 4-2 Country rock $\delta^{18}O$ results. Isotopic results for the same five units as in figure 4-1. Dot - silicates (this study), open circles - gneisses from the Prince Rupert area (Magaritz & Taylor, 1976b), C - carbonates. Horizontal bars are the granitoid rock range for the respective areas.

granites found in the Coast Mountains (Hutchison, 1970, Roddick & Hutchison, 1974) and have been shown to be geochemically very similar to the calc-alkaline intrusives in the belt (Barker & Arth, 1982). The primitive $\delta^{18}O$ character of the Coast Range batholiths is compatible with their derivation from these relatively low-^{18}O materials. The three samples of Kluane Schist are slightly richer in ^{18}O than the granulitic rocks discussed above, and could be responsible for the observed increase in the $^{18}O/^{16}O$ ratios of the quartz separates in the quartz diorites of the Skagway traverse and in the granodioritic samples of this study (Mgd-3, -1, -4).

The Whitehorse Trough country rocks are the most extensively sampled in the study and exhibit the largest variability in their $^{18}O/^{16}O$ ratios, most of which is due to lowering by hydrothermal activity. Basic to intermediate volcanics (four samples) range from 7.5 to 11.2 values compatible with sea-water weathering at low temperatures in a fore-arc or back-arc basinal environment (Tempelman-Kluit, 1979, Bultman, 1979). This type of alteration has been extensively studied in MORBs and in ophiolitic sections (Garlick & Dymond, 1970, Muehlenbachs & Clayton, 1972, Gregory & Taylor 1981) and tends to increase the $\delta^{18}O$ values of basalts up to 10°/oo.

Only two samples of unaltered clastic sedimentary rocks (arkosic sandstone and siltstone) have been analysed, yielding results within the normal +9 to +16 range for these

rock types (Savin & Epstein, 1970b).

Although the primary ^{18}O of granitoid rocks in the Trough is similar in composition to the values observed in basic volcanics in the region, there does not seem to be a direct association between batholiths and surounding rocks in this terrane. This is clearly indicated by the large spread in values for the country rocks and the restricted range observed in granitoid rocks.

Finally in the Omineca Belt, metasedimentary rocks are virtually identical in $\delta^{18}O$ (+9.6 to +17.6°/₀₀) to the granites cropping out in this area. Some of the lowest values, found in metavolcanic schists and volcanic rocks, occur along the margins of the Omineca Belt, in the Teslin Suture Zone, and may have been depleted by interaction with meteoric water during intrusive periods in the Cretaceous. Country rocks of the type seen in the area but buried deeper in the crust, could have produced the ^{18}O compositions found in the Omineca quartz monzonites.

C. Comparison of ^{18}O and Strontium isotopes

The use of strontium combined with oxygen isotope data has become widespread in recent years in interpreting the sources of magmas. Studies tend to either concentrate on a single plutonic complex using a variaty of techniques or to look at regional trends in ^{18}O and $^{87}Sr/^{86}Sr$ variations. In the latter case, only broad isotopic and chemical characteristics can be deduced about the presumed sources

and the conclusions are strongly based on theoretical models (Harmon & Halliday, 1980, Haack *et al.*, 1982, Magaritz *et al.*, 1978, Longstaffe *et al.*, 1983)

Because of the problem of hydrothermal alteration affecting the $^{18}O/^{16}O$ ratios of many rocks of this study on which strontium data was available, only a very small number (about 20) are suitable for comparison of oxygen and strontium. Figure 4-3 is a compilation of Sr data from LeCouteur and Tempelman-Kluit (1976), Morrison and others (1979) and R. Armstrong (unpublished data) for rocks in the normal ^{18}O range (filled symbols) and depleted rocks (open symbols). No correlation exists between the initial $^{87}Sr/^{86}Sr$ ratio and Sr content or $\delta^{18}O$ for the granitoids of the three western terranes. Initial Sr variation seems to be more related to geography than to surface geology (cf. references above).

The initial $^{87}Sr/^{86}Sr$ of the Yukon Crystalline Belt granites range from 0.7061 to 0.7070; values thought to be indicative of the presence of underlying Precambrian continental crust (Kistler & Peterman, 1973, Armstrong *et al.*, 1977). A general decrease in Sr initial ratios is observed from the northernmost rocks to the southwestern parts of the terrane. In the eastern part of the Coast Mountains and in the Whitehorse area, unaltered granitoids have initial ratios varying from 0.7041 to 0.7054, but most plot within the 0.704 to 0.705 ratio range. This is within the inferred present-day mantle range of 0.702 to 0.705

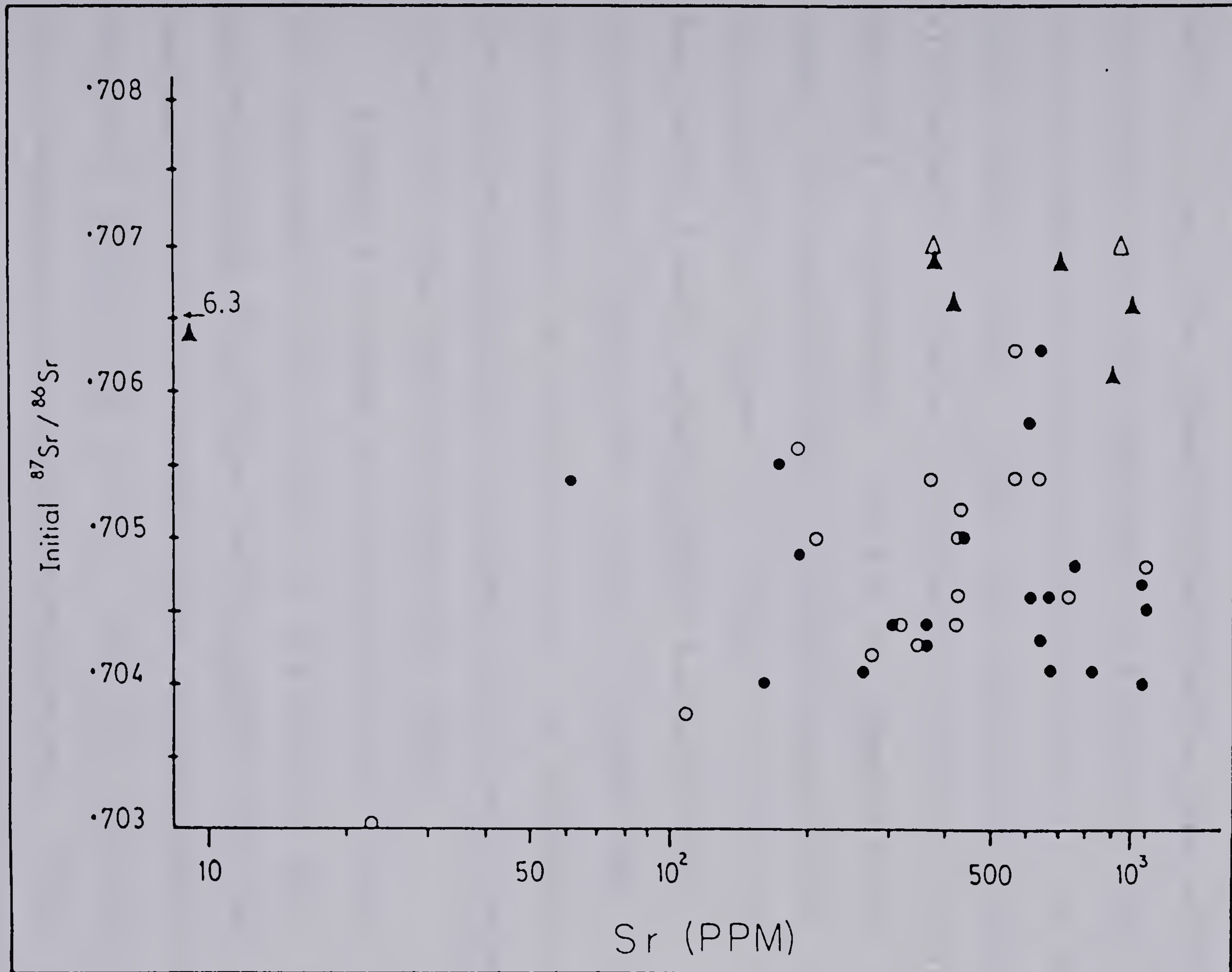


Figure 4-3 <u>Initial $^{87}Sr/^{86}Sr$ ratios versus Sr content in the granitoid rocks</u> of the YCT (triangles), and the WT+CPC (dots). Open symbols represent altered rocks of the respective areas.

(Hofmann & Hart, 1978). Altered rocks (open circles) plot slightly higher than unaltered samples on figure 4-3; it is not known whether this is a causal relationship or not.

The lack of correlation between Sr content and initial ratios in both the Yukon Crystalline Terrane and the Whitehorse and Coast provinces is similar to what is found in calc-alkaline volcanics and intrusives in the Central Andes (McNutt *et al.*, 1975). The trend is not compatible with models of crustal assimilation since the crust is enriched in radiogenic ^{87}Sr but has generally low ($\leq$100 ppm) common Sr concentrations. A crustal assimilation trend would be expected to increase the initial $^{87}Sr/^{86}Sr$ of Sr-poor samples to a much larger extent than of Sr-rich granodioritic batholiths. Figure 4-3 shows that, through the entire range of Sr concentrations, the radiogenic content of the samples stays about constant within each respective group (*i.e.* the [YCT] and the [WT + CPC]).

Figure 4-4 shows the relationship of $\delta^{18}O$ to $^{87}Sr/^{86}Sr$ for the rocks of this study. In the Omineca Belt, values are typically high for oxygen, which agrees with their postulated anatectic origin based on previous Sr isotope work that showed high initial $^{87}Sr/^{86}Sr$ (Fairbairn *et al.*, 1964, Wanless *et al.*, 1968, Godwin *et al.*, 1980). Both oxygen and strontium results in the plutonic samples are typical of ratios obtained in older (Precambrian or Lower Paleozoic) sialic crustal components. If these rocks were not produced directly from old crustal material then, at the

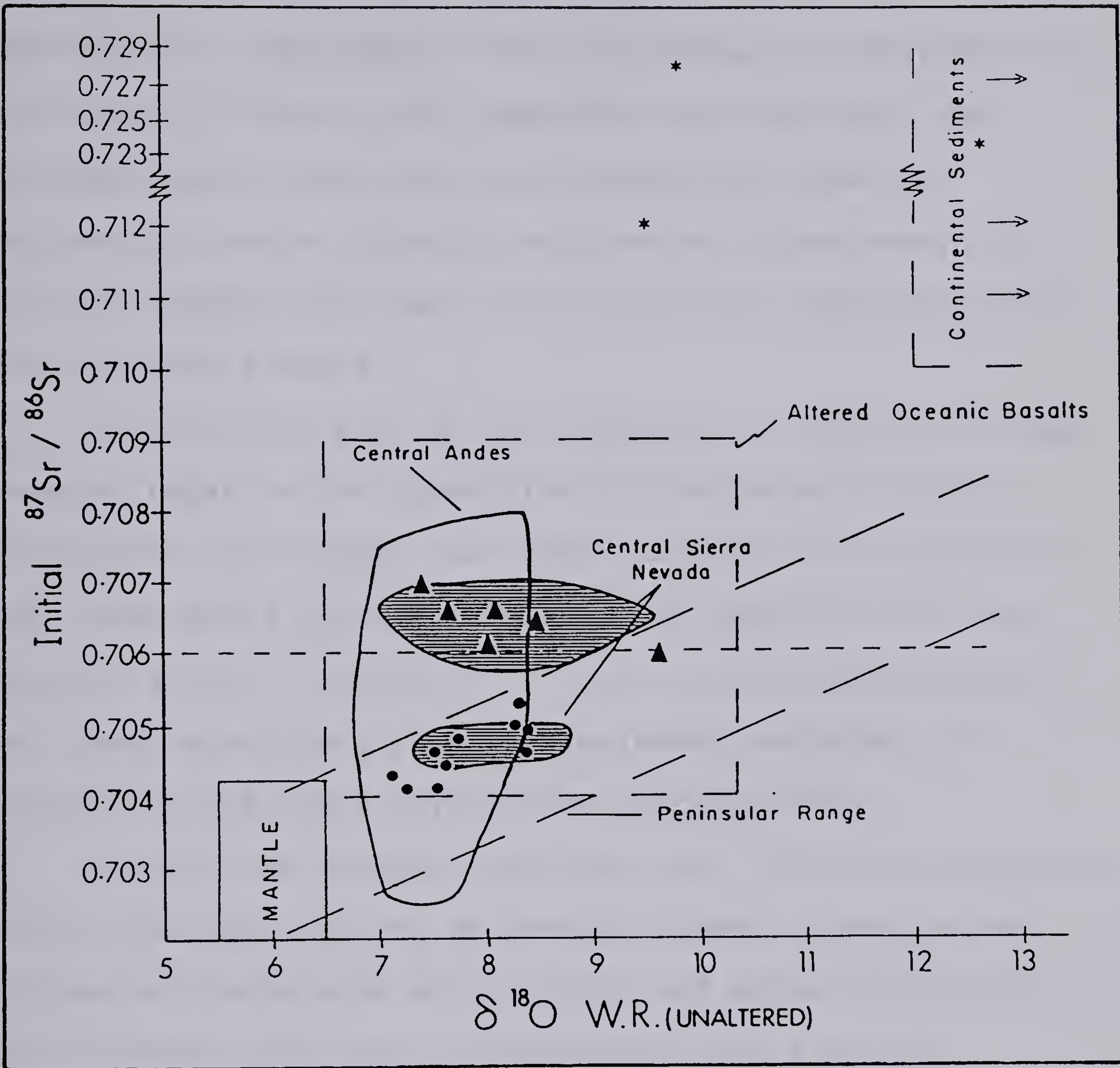


Figure 4-4 <u>Initial $^{87}Sr/^{86}Sr$ ratios versus $\delta^{18}O$ of rocks from the Yukon.</u> Dots are the Whitehorse Trough and the Coast Range rocks; triangles, Yukon Crystalline Terrane; and *, the eastern rocks. The dashed trend is from the Peninsular Range batholith in California (Taylor & Silver, 1978). The Central Andean range comes from Longstaffe and others (1983); the Sierra Nevada from Masi and others (1981). The mantle box was defined relative to the Sr range of Hofmann and Hart (1978) and to the oxygen isotope range of fresh MORBs (Muehlenbachs and Clayton, 1972).

very least, they must have interacted with that reservoir during their emplacement. Strontium data is unavailable for the country rocks in the immediate area but their age (Precambrian to Paleozoic) and composition (marine sedimentary rocks including pelites and limestones, and volcanic rocks) and their $\delta^{18}O$ values are compatible with the hypothesis above.

In the lower part of the diagram, two distinct groups emerge; rocks in the Yukon Crystalline Terrane (filled triangles) plot higher than those from the Whitehorse and the Coast belts (filled circles). The separation of the groups is not a function of ^{18}O as was discussed earlier, but results entirely from the increased radiogenic Sr content of the Yukon Crystalline Terrane rocks.

Most of the samples from the lower $^{87}Sr/^{86}Sr$ group plot within the area defined by Central Andean volcanism and plutonism (Longstaffe et al, 1983) and agree fairly well with results obtained in the Central Sierra Nevada batholiths (Masi *et al.*, 1981). These granitoid rocks have both primitive oxygen and unradiogenic Sr signatures and require little or no involvement of upper crustal material to generate the observed trend. Other granitoid plutons in the Intermontane Belt such as the Mesozoic Guichon Creek Batholith and surrounding Nicola Group volcanics also have low initial $^{87}Sr/^{86}Sr$ (0.7034) (Chrismas *et al.*, 1968, Preto *et al.*, 1979) and normal "I" type ^{18}O results (Taylor and Magaritz , 1978). The calc-alkaline nature of the

Intermontane Belt granitoids and associated volcanics and the above isotopic characteristics suggest a deep seated, arc-related origin for these magmas. Morrison and others (1979) proposed that the Whitehorse Trough was underlain by Paleozoic oceanic crust through which the intrusives must have passed during their ascent to upper crustal levels.

The top layer of tholeiitic oceanic crust becomes progressively enriched in ^{18}O and ^{87}Sr by interaction with seawater ($\delta^{18}O$ = 0‰, $^{87}Sr/^{86}Sr \cong 0.709$) at low temperature (Muehlenbachs & Clayton, 1972, Staudigel *et al.*, 1981, McCulloch *et al.*, 1980). It is possible that the small positive slope observed in figure 4-4 for the rocks of the eastern Coast Range rocks and the Whitehorse Trough was caused by variable amounts of assimilated altered and pristine oceanic crust, but the intrinsic variability of such rocks, in both oxygen and strontium, does not allow, at present, to clearly define the extent of this process on the observed isotopic compositions.

The Yukon Crystalline Terrane is more difficult to interpret. The position it occupies on figure 4-4, shows primitive ^{18}O and higher than normal radiogenic Sr. If these granitoids are also arc-related as has been suggested by Gabrielse & Reesor (1974) and Tempelman-Kluit (1976, 1979) then source contamination from subducted marine sediments could be an explanation for the enriched radiogenic character of these rocks. Continent-derived sediments are richer in both ^{18}O and radiogenic Sr (Savin & Epstein,

1970a, Dasch, 1969) than the basaltic ocean floor they cover. James (1980) suggested that subduction of marine sediments along with the oceanic crust, would increase the amounts of incompatible elements such as Rb and Sr in melts derived from the subducted slab over the concentrations found in the mantle. This would be expected if preferential melting of the sediments occurred because of increased temperatures at the top of the slab, as suggested by Oxburgh and Turcotte (1970). Since both Sr contents and $^{87}Sr/^{86}Sr$ ratios are much larger in the subducted sediments and the slab rocks than in the surrounding mantle wedge, the magmas derived from melting of these materials would be strongly buffered by the crustal component of strontium. The $^{18}O/^{16}O$ ratios though, should not behave similarly, because the large reservoir of oxygen in the mantle would limit any enrichment in the magmas to less than about 1-2‰. A process such as this could produce the pattern shown by the Yukon Crystalline Terrane rocks.

Experimental studies though, have shown that melts of granodioritic and quartz monzonitic nature cannot be directly generated at mantle pressures and temperatures (Wyllie *et al.*, 1976, Stern *et al.*, 1975) because of the presence of quartz on the liquidus in melts of such composition. This would require that slab magmas undergo chemical (and isotopic?) transformations at higher levels in the lithosphere prior to their final emplacement as granodioritic plutons in the continental crust. Also, it is

not clear if the subducted marine sediments necessary in James's model would persist at the depths required for calc-alkaline magmatism (≃150-200 km). Huang and Wyllie (1973) have indicated that this type of material should not persist in its original hydrous form much beyond 30 km depth. Conversely, studies of andesitic volcanism in the Banda Arc (Magaritz *et al.*, 1978) have shown James's "characteristic" source contamination trend (*i.e.* small increase in ^{18}O and comparatively larger increase in radiogenic ^{87}Sr) and therefore a similar process cannot be rejected in the case of the Yukon Crystalline Terrane intrusives. The evidence though, is inconclusive, at present, because of the problems associated with this model.

Another possible model includes magma generation in the lower crust. As mentioned earlier, highly metamorphosed rocks from cratonic environments tend to have relatively low ^{18}O concentrations (+5 to +9°/₀₀). This is true for granulitic rocks of sedimentary and igneous origin and for metasedimetary and metaigneous high-grade rocks in the Canadian Shield. A process of interaction with a deep mantle-like oxygen reservoir was suggested in these rocks to explain their depletion in ^{18}O (Fourcade & Javoy, 1973, Shieh & Schwarcz, 1974). Alternatively, it is possible that the low-^{18}O character of Archean high-grade metamorphic rocks is a primary feature of these rocks. This could be due to higher temperatures of erosion and weathering during the early history of the earth or possibly because Archean

sialic rocks were juvenile additions to the continental crust (Longstaffe, 1979, Longstaffe & Birk, 1981, Longstaffe & Gower, 1983). If the Yukon Crystalline Terrane granitoids were derived from melting of very old high-grade metamorphic rocks, then their final $\delta^{18}O$'s should not be significantly different from what is observed at present. The required radiogenic ^{87}Sr component of the plutons could be satisfied by the addition of variable amounts of phases such as biotite and clay minerals during their ascent through the upper crust.

Low ^{18}O high-grade gneisses do crop out in the Prince Rupert area (Magaritz & Taylor, 1976b) but their unradiogenic nature (0.704 to 0.7057) is not compatible with the ratios observed in the Yukon Crystalline Terrane granitoids (Armstrong & Runkle, 1979). Metamorphic rocks of appropriate $\delta^{18}O$ composition do occur in the Yukon Crystalline Terrane (figure 4-2) but their strontium isotopic content is as yet unknown. Because of the many variables and assumptions involved in present magma genesis models, the choice between source contamination or lower crustal generation and crustal contamination cannot be made, based on the limited data available for the Yukon Crystalline Terrane granitoid rocks. Future lead isotope and trace element studies should resolve this issue. At present, the fact that upper crustal oxygen has not interacted with the batholiths is unambiguous.

D. Summary

^{18}O and radiogenic isotope data of the granitoids of the study area show the following characteristics;

1. The Whitehorse Trough plutons have primitive $\delta^{18}O$ and relatively unradiogenic $^{87}Sr/^{86}Sr$ ratios implying deep seated sources for the calc-alkaline rocks. Upper crustal isotopic exchange with continental rocks at magmatic temperatures, if present, is minimal in the area. Alternatively, ^{18}O exchange with variably altered oceanic crust and arc-derived volcanic rocks during emplacement of the plutons of the Whitehorse Trough is a realistic possibility.
2. The eastern margin rocks of the Coast Mountains have isotopic features that renders them indistinguishable from plutons of the Whitehorse Trough. Western margin CPC rocks have a slightly enriched ^{18}O content, that may have resulted from interaction with the surrounding country rocks.
3. The Yukon Crystalline Terrane granitoid rocks are similar in $\delta^{18}O$ to rocks from the two previous terranes but are enriched in radiogenic ^{87}Sr. They could be derived from melting of oceanic sedimentary material in a subduction environment or alternatively, could have been produced in the deep continental crust and contaminated by crustal radiogenic Sr from pelitic and gneissic rocks.
4. Atlin Terrane and Omineca Belt granites have high $\delta^{18}O$

signatures, while the Omineca Belt granites also have very radiogenic strontium. Large scale interaction with upper crustal sedimentary oxygen and strontium has definitely occurred in these rocks. The $\delta^{18}O$ values of Atlin granites seem to be affected by the surrounding country rocks to a large extent; high $\delta^{18}O$ rocks crop out in marine sedimentary piles while lower $\delta^{18}O$ plutons intrude basic and ultrabasic volcanic rocks.

Conclusion

Oxygen isotope analysis of granitoid rocks from the southern and central Yukon shows two distinct groups of plutons; (1) A low $\delta^{18}O$ (+6 to 8.4°/oo), suite of large batholiths of quartz dioritic to granodioritic to quartz monzonitic composition which were emplaced in volcanic, volcaniclastic and crystalline terranes, and (2) A high $\delta^{18}O$ group (+8 to +12°/oo) of compositionally restricted granites and quartz monzonites that intruded marine sedimentary and metasedimentary rocks.

Rocks from the Whitehorse Trough, the Coast Plutonic Belt and the Yukon Crystalline Terrane have similar primitive $\delta^{18}O$ compositions indicating that upper crustal oxygen has not played a significant part in their formation. Strontium isotopes suggest that the Yukon Crystalline Terrane granitoid rocks may have interacted with lower crustal (Precambrian basement?) radiogenic material or that source contamination caused the observed increase in ^{87}Sr. The Omineca Belt granites exhibit high $\delta^{18}O$ and initial $^{87}Sr/^{86}Sr$ ratios typical of rocks derived directly from crustal sources.

Hydrothermal alteration has been pervasive in locations where country rocks are permeable to groundwater flow. This is especially true of the Whitehorse batholiths and stocks which are petrographically and isotopically altered by meteoric water down to whole-rock $\delta^{18}O$ values of -1°/oo. Water-rock ratios in the granites of the area are variable,

the highest being about 0.9; a value similar to those found in other volcanic areas intruded by epizonal plutons. Much higher water fluxes (W/R >>1) through mineralised skarns and metapelites has locally produced rocks in the -7 to -10°/₀₀ range and even lower for vein minerals. Temperatures of alteration were moderate to low (400-200°C), based on mineral assemblages.

Similar hydrothermal alteration patterns in the Atlin Terrane and Omineca Belt granites did not produce depleted $\delta^{18}O$ signatures. Fluids circulating in those areas were dominantly magmatic or ^{18}O shifted groundwaters. In the latter case, the shift toward more enriched $\delta^{18}O$ values for the water must have been caused through interaction with high ^{18}O metasedimentary rocks cropping out in the area.

Two main hydrothermal events have been outlined, one in Middle Cretaceous time (110 to 90 Ma) and one in Eocene time (65 to 50 Ma) The latter event has been associated with the great majority of porphyry copper deposits in Western North America (Godwin, 1975, Titley & Beane, 1981). K-Ar ages of altered rocks may have been reset by hydrothermal heating and alteration during the periods of climactic plutonism. Rb-Sr dating of the propylitised samples usually yields poor results because of isotope heterogeneities which may or may not have been caused by the alteration event. Future stable and radiogenic isotope studies concerned with primary features should concentrate sampling away from contacts with permeable country rocks, preferably on coarser grained

rock-types, as they are less likely to be altered.

The pervasive character of hydrothermal alteration in the study area, should not be considered to be the result of sampling biases toward altered rocks but a primary feature of many batholiths in the northern Canadian Cordillera. Similar alteration periods must have occurred throughout many other regions in North America but the high latitude position of the Yukon granites makes them especially amenable to the study of fossil hydrothermal systems.

References cited

Anderson, A.T., Jr., Clayton, R.N., and Mayeda, T.K. 1971. Oxygen isotope thermometry of mafic igneous rocks: *Jour. Geol.* **79**, p. 715-729.

Anderson, R.G. 1983a. Potential of SE Selwyn plutonic suite for W-skarn deposits. (Abs.) *Geol. Assoc. Can. & Mineral. Assoc. Can. Progr. with Abs.* **8**, p. A2.

Anderson, R.G. 1983b. Two different granitoid types in the Northern Canadian Cordillera and comparisons with other fold belts. (Abs.) *Geol. Assoc. Can. & Mineral. Assoc. Can. Progr. with Abs.* **8**, p. A2.

Armstrong, R.L. 1974. Geochronometry of the Eocene volcanic-plutonic episode in Idaho: *Northwest Geology* **3**, p. 1-15.

Armstrong, R.L. 1975. The geochronometry of Idaho. *Isochron/West* **14**, 50p.

Armstrong, R.L., and Runkle, D.E. 1979. Rb-Sr geochronometry of Ecstall, Kitkiata, and Quottoon plutons and their country rocks, Prince Rupert region, Coast Plutonic Complex, British Columbia. *Can. Jour. Earth Sci.* **16**, 387-399.

Armstrong, R.L., Taubeneck, W.H., and Hales, P.O. 1977. Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington and Idaho. *Geol. Soc. Am. Bull.* **88**, p. 397-411.

Barker, F., and Arth, J.G. 1982. "Central Gneiss Complex" of the Coast Range Batholith: the roots of a high-K, calc-alkaline arc ? (Abs.) *Geol. Soc. Am. Abs. & Prog.* **14** No 7, p. 439.

Beane, R.E., and Titley, S.R. 1981. Porphyry copper deposits. Part II: Hydrothermal alteration and mineralization. *Econ. Geol. 75th Anniversary Volume*, p. 235-269.

Blattner, P. and Bird, G.W. 1974. Oxygen isotope fractionation between quartz and K-feldspar at 600 celsius. *Earth Planet. Sci. Lett.* **23**, p. 21-27.

Blattner, P., Dietrich, V., and Gansser, A. 1983. Contrasting ^{18}O enrichment and origins of High Himalayan and Transhimalayan intrusives. *Earth Planet. Sci. Lett.* **65**, p. 276-286.

Bottinga, Y. 1968. Calculation of fractionation factors for carbon and oxygen exchange in the system calcite-carbon dioxide-water: *Jour. Phys. Chem.* **72**, p. 800-808.

Bottinga, Y., and Javoy, M. 1975. Oxygen isotope partitioning among the minerals in igneous and metamorphic rocks. *Rev. Geophys. Space Phys.* **13**, p. 401-418.

Bowman, J.R., and Covent, J. 1983. Hydrogen, carbon, and oxygen isotope studies of the Cantung scheelite skarn, Tungsten, N.W.T., Canada. (Abs.) *Geol. Assoc. Can. & Mineral. Assoc. Can. Prog. with Abs.* **8**, p. A7.

Bultman, T.R. 1979. Geology and tectonic history of the Whitehorse Trough west of Atlin, British Columbia. Unpub. Ph.D. thesis. Yale University. New Haven. CT. 284p.

Cathles, L.M. 1977. An analysis of cooling intrusives by groundwater convection which includes boiling. *Econ. Geol.* **72**, p. 804-826.

Chappell, B.W., and White, A.J.R. 1974. Two contrasting granite types. Pacific Geol. 8, p. 173-174.

Chivas, A.R., Andrews, A.S., Sinha, A.K., and O'Neil, J.R. 1982. Geochemistry of a Pliocene-Pleistocene oceanic-arc plutonic complex, Guadalcanal. *Nature* **300**, p. 139-143.

Chrismas, L., Baadsgaard, H., Folinsbee, R.E., Fritz, P., Krouse, H.R., and Sasaki, A. 1968. Rb/Sr, S and O isotopic analyses indicating source and date of contact metasomatic copper deposits, Craigmont, B.C., *Econ. Geol.* **64**, p. 479-488.

Clauer, N. 1980. Strontium and argon isotopes in naturally weathered biotites, muscovites and feldspars. *Chem. Geol.* **31**, p. 325-334.

Clauer, N., O'Neil, J.R., and Bonnot-Courtois, C. 1982. The effect of natural weathering on the chemical and isotopic composition of biotites. *Geochim. Cosmochim. Acta.* **46**, p. 1755-1762.

Clayton, R.N., and Mayeda, T.K. 1963. The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis: *Geochim. Cosmochim. Acta* **27**, p. 43-52.

Clayton, R.N., O'Neil, J.R., and Mayeda, T. 1972. Oxygen

isotope exchange between quartz and water. *Jour. Geophys. Res.* 77, p. 3057-3067.

Cole, D.R., and Ohmoto, H. 1976. Effect of NaCl on the rate of oxygen isotopic exchange reactions between rocks and water. (Abs.) *Geol. Soc. Am. Abs. & Prog.* 8 No 6, p. 817.

Cole, D.R., and Ohmoto, H. 1979. Mechanisms of oxygen isotope exchange between aqueous solutions and granitic rocks. (Abs.) *Geol. Soc. Am. Abs. with Prog.* 11 No 7, p. 403.

Craig, H. 1957. Isotopic standards for carbon and oxygen and correction factors for mass-spectrometric analysis of carbon dioxide. *Geochim. Cosmochim. Acta.* **12**, p. 133-149.

Craig, H. 1961a. Standard for reporting concentrations of deuterium and oxygen-18 in natural waters. *Science* **133**, p. 1833-1834.

Craig, H. 1961b. Isotopic variations in meteoric waters. *Science* **133**, p. 1702-1703.

Craig, H. 1963. The isotope geochemistry of water and carbon in geothermal areas. *in* E. Tongiori, ed., Nuclear Geology on Geothermal Areas, Spoleto. CNR, Lab. Geol. Nucl. Pisa., p. 17-53.

Criss, R.E., Lanphere, M.A., and Taylor, H.P. Jr. 1982. Effects of regional uplift, deformation, and meteoric-hydrothermal metamorphism on K-Ar ages of biotites in the southern half of the Idaho batholith. *Jour. Geophys. Res.* 87, p. 7029-7046.

Criss, R.E., and Taylor, H.P. Jr. 1982. An $^{18}O/^{16}O$ and D/H study of Tertiary hydrothermal systems in the southern half of the Idaho batholith. *Geol. Soc. Am. Bull.* **94**, p.640-663.

Dagenais, G.R. 1983. Alteration of Whitehorse Trough granitoids as detected by $\delta^{18}O$; Its implication for Rb-Sr geochronology. (Abs.) *Geol. Assoc. Can. & Mineral. Assoc. Can. Progr. with Abs.* 8, p. A15.

Dansgaard, W. 1964. Stable isotopes in precipitation. *Tellus* **16**, p. 436-468.

Dasch, E.J. 1969. Strontium isotopes in weathering profiles, deep sea sediments, and sedimentary rocks. *Geochim. Cosmochim. Acta* **33**, p. 1521-1552.

Degens, E.T., and Epstein, S. 1962. Relationship between

$^{18}O/^{16}O$ ratios in coexisting carbonates, cherts, and diatomites. *Am. Assoc. Petrol. Geol. Bull.* **46**, p. 534-542.

Degens, E.T., and Epstein, S. 1964. Oxygen and carbon isotope ratios in coexisting calcites and dolomites from recent and ancient sediments. *Geochim. Cosmochim. Acta.* **28**, p. 23-44.

Deines, P. 1977. On the oxygen isotope distribution among mineral triplets in igneous and metamorphic rocks. *Geochim. Cosmochim. Acta* **41**, p. 1709-1730.

Deines, P., and Gold, D.P. 1969. The change in carbon and oxygen isotopic composition during contact metamorphism of Trenton limestone by the Mount Royal pluton. *Geochim. Cosmochim. Acta* **33**, p. 421-424.

Epstein, S., and Mayeda, T. 1953. Variation of O^{18} content of waters from natural sources. *Geochim. Cosmochim. Acta.* **4**, p. 213-224.

Epstein, S., and Taylor, H.P., Jr. 1971. $^{18}O/^{16}O$, Si^{30}/Si^{28}, D/H, and $^{13}C/^{12}C$ ratios in lunar samples, *Proc. Second Lunar Sci. Conf.* **2**, Cambridge Mass., M.I.T. Press, p. 1421-1441.

Fairbairn, H.W., Hurley, P.H., and Pinson, W.K. 1964. Initial $^{87}Sr/^{86}Sr$ and possible sources of granitic rocks, southwest British Columbia. *Jour. Geophys. Res.* **69**, p. 4889-4893.

Forester, R.W., and Taylor, H.P. Jr 1972. Oxygen and hydrogen isotope data on the interaction of meteoric groundwaters with a gabbro-diorite stock, San Juan Mountains, Colorado. *24th Internat. Geol. Cong.*, Montreal, sec. 10, Geochemistry, p. 254-263.

Forester, R.W., and Taylor, H.P., Jr. 1974. Oxygen, hydrogen and carbon isotope systematics in the Early Tertiary igneous province of Skye, northwest Scotland. *Geol. Soc. Am. Abstr. Prog.* **6**, p. 742.

Forester, R.W., and Taylor, H.P. Jr. 1977. $^{18}O/^{16}O$, D/H and $^{13}C/^{12}C$ studies of the Tertiary igneous complex of Skye, Scotland. *Amer. Jour. Sci.* **277**, p. 136-177.

Fourcade, S., and Javoy, M. 1973. Rapports $^{18}O/^{16}O$ dans les roches du vieux socle catazonal d'In Ouzzal (Sahara Algérien). *Contrib. Mineral. Petrol.* **42**, p. 235-244.

Friedman, I., O'Neil, J.R. 1977. Compilation of stable isotope fractionation factors of geochemical interest. *in* Data of Chemistry, Chapter KK, M.

Fleischer, ed., *U. S. Geol. Surv. Prof. Pap.* **440-KK**, 12 p., 48 figs., 6th ed.

Friedman, I., Redfield, A.C., Schoen, B., and Harris, J. 1964. The variation in the deuterium content of natural waters in the hydrologic cycle. *Rev. Geophys.* **2**, p. 177-224.

Gabrielse, H., and Reesor, J.E. 1974. The nature and setting of granitic plutons in the central and eastern parts of the Canadian Cordillera. *Pacific Geol.* **8**, p. 109-138.

Gabrielse, H. and Wheeler, J.O. 1961. Tectonic framework of southern Yukon and northern British Columbia. *Geol. Surv. Can. Paper* **60-24**, 37p

Garlick, G.D., and Dymond, J.R. 1970. Oxygen isotope exchange between volcanic materials and ocean water. *Geol. Soc. Am. Bull.* **81**, p. 2137-2142.

Garlick, G.D., and Epstein, S. 1967. Oxygen isotope ratios in coexisting minerals of regionally metamorphosed rocks. *Geochim. Cosmochim. Acta* **31**, p. 181-214.

Giletti, B.J., and Nagy, K.L. 1981. Grain boundary diffusion of oxygen along lamellar boundaries in perthitic feldspar.(Abs.) *EOS Trans. Am. Geophys. Union* **62**, p.428.

Godwin, C.I. 1975. Imbricate subduction zones and their relationship with Upper Cretaceous to Tertiary porphyry deposits in the Canadian Corillera: *Can. Jour. Earth Sci.* **12**, p. 1362-1378.

Godwin, C.I., Armstrong, R.L., and Tompson, K.M. 1980. K-Ar and Rb-Sr dating and the genesis of tungsten at the Clea tungsten skarn property, Selwyn Mountains, Yukon Territory. *CIM Bulletin*, p. 90-93.

Goodfellow, W.D., Jonasson, I.R., Smee, B.W. 1978. Uranium in Nisling Range alaskite and related rocks of the Yukon Crystalline Terrane : Discussion. *Geol. Surv. Can. Paper* **78-1A**, p.555-556.

Green, T.H. 1976. Experimental generation of cordierite- or garnet-bearing granitic liquids from a pelitic composition, *Geology* **4**, p. 85-88.

Gregory, R.T. and Taylor, H.P. Jr. 1981. An oxygen isotope profile in a section of Cretaceous oceanic crust, Samail ophiolite, Oman: evidence for ^{18}O-buffering of the oceans by deep (>5km) seawater-hydrothermal circulation at mid-ocean ridges. *Jour. Geophys. Res.*

86, p. 2737-2755

Haack, U., Hoefs, J., and Gohn, E. 1982. Constraints on the origin of Damaran granites by Rb/Sr and $\delta^{18}O$ data. *Contrib. Mineral. Petrol.* **79**, p. 279-289.

Harmon, R.S., and Halliday, A.N. 1980. Oxygen and strontium isotope relationships in the British Late Caledonian granites. *Nature* **283**, p. 21-25.

Harrison, T.M., Armstrong, R.L., Naeser, C.W., and Harakal, J.E. 1979. Geochronology and thermal history of the Coast Plutonic Complex, near Prince Rupert, British Columbia. *Can. Jour. Earth Sci.* **16**, p. 400-410.

Hattori, K., and Muehlenbachs, K. 1982. Oxygen isotope ratios of the Iclandic crust. *Jour. Geophys. Res.* 87, p.6559-6565.

Haul, R., and Dümbgen, G. 1962. Untersuchung der Sauerstoffbeweglichkeit in Titandioxid, Quartz und Quartzglass mit Hilfe des heterogenen Isotopenaustausches. *Zeits. Elektrochem.* **66**, p. 636-641.

Haynes, F.M., and Titley, S.R. 1980. The evolution of fracture-related permeability within the Ruby Star granodiorite, Sierrita porphyry copper deposits, Pima County, Arizona. *Econ. Geol.* **75**, p. 673-683.

Hofmann, A.W., Hart, S.R. 1978. An assessment of local and regional isotopic equilibrium in the mantle. *Earth Planet. Sci. Lett.* **38**, p. 44-62.

Huang, W.L., and Wyllie, P.J. 1973. Muscovite dehydration and melting in the deep crust and subducted ocean sediments. *Earth Planet. Sci. Lett.* **18**, p. 133-136.

Hutchison, W.W. 1970. Metamorphic framework and plutonic styles in the Prince Rupert region of central British Columbia. *Can. Jour. Earth Sci.* **7**, p. 376-405.

Ishihara, S. 1975. The magnetite-series and ilmentite-series granitic rocks. *Mining Geol.* **27**, p. 293-305.

Ito E., and Clayton R.N. 1983. Submarine metamorphism of gabbros from the Mid-Cayman Rise: an oxygen isotope study. *Geochim. Cosmochim. Acta* **47**, p. 535-546.

James, D.E. 1981. The combined use of oxygen and radiogenic isotopes as indicators of crustal contamination. *Ann. Rev. Earth Planet. Sci.* **9**, p. 311-344.

Javoy, M. 1977. Stable isotopes and geothermometry. *Jour Geol. Soc. Lond.* **133**, p. 609-636.

Keith, M.L., and Weber, J.N. 1964. Carbon and oxygen isotopic composition of selected limestones and fossils. *Geochim. Cosmochim. Acta* **28**, p. 1787-1816.

Kistler, R.W., and Peterman, Z.E. 1973. Variations in Sr, Rb, K, Na and initial $^{87}Sr/^{86}Sr$ in Mesozoic granitic rocks and intruded wall rocks in central California. *Geol. Soc. Amer. Bull.* **84**, p. 3489-3512.

Knauth, L.P., and Epstein, S. 1976. Hydrogen and oxygen isotope ratios in nodular and bedded cherts. *Geochim. Cosmochim. Acta.* **40**, p. 1095-1108.

Knauth, L.P., and Lowe, D.R. 1978. Oxygen isotope geochemistry of cherts from the Onverwacht group (3.4 billion years) Transvaal, South Africa, with implication for secular variations in the isotopic composition of cherts. *Earth Planet. Sci. Lett.* **41**, p. 209-222.

Lambert, M.B. 1974. The Bennet Lake cauldron subsidence complex, British Columbia and Yukon Territory. *Geol. Surv. Can. Bull.* **227**, 213p.

LeCouteur, P.C. and Tempelman-Kluit, D.J. 1976. Rb/Sr ages and a profile of initial $^{87}Sr/^{86}Sr$ ratios for plutonic rocks across the Yukon Crystalline Terrane. *Can. Jour. Earth Sci.* **13**, p. 319-330.

Longstaffe, F.J. 1979. The oxygen-isotope geochemistry of Archean granitoids, *in* F. Barker, Ed., Trondhjemites and related rocks. Elsevier, Amsterdam, p. 363-399.

Longstaffe, F.J. 1982. Stable isotopes in the study of granitic pegmatites and related rocks. *in* P. Cerny, Ed., Granitic pegmatites in science and industry. Min. Assoc. Can. short course handbook 8, Toronto, Ontario, p. 373-404.

Longstaffe, F.J., and Birk, D. 1981. Oxygen isotope evidence for the origin of discrete granitoid plutons from the Archean Western Wabigoon Belt, northwestern Ontario. *Precamb. Res.* **14**, p. 21-30.

Longstaffe, F.J., Cerny, P., and Muehlenbachs, K. 1981. Oxygen-isotope geochemistry of the granitoid rocks in the Winnipeg River Pegmatite District, southeastern Manitoba. *Can. Mineral.* **19**, p. 195-204.

Longstaffe, F.J., Clark, A.H., McNutt, R.H., and Zentilli, M. 1983. Oxygen isotopic compositions of Central

Andean plutonic and volcanic rocks, latitudes 26° - 29° south. *Earth Planet. Sci. Lett.* **64**, p. 9-18.

Longstaffe, F.J., and Gower, C.F. 1983. Oxygen isotope geochemistry of Archean granitoid gneisses and related rocks in the English River subprovince, northwestern Ontario. *Precamb. Res.* **22**, p. 203-218.

Longstaffe, F.J., Nesbitt, B.E., and Muehlenbachs, K. 1982. Oxygen isotope geochemistry of the shales hosting Pb-Zn-Ba mineralization at the Jason Prospect, Selwyn Basin, Yukon. *Geol. surv. Can. Paper* 82-1C, p. 45-49.

Longstaffe, F.J., and Schwarcz, H.P. 1977. $^{18}O/^{16}O$ of Archean clastic metasedimentary rocks: a petrogenetic indicator for Archean gneisses? *Geochim. Cosmochim. Acta* **41**, p. 1303-1312.

Longstaffe, F.J., Smith, T.E., and Muehlenbachs, K. 1980. Oxygen isotope evidence for the genesis of Upper Paleozoic granitoids from southwestern Nova Scotia. *Can. Jour. Earth Sci.* **17**, p. 132-141.

Lowdon, J.A. 1961. Age determinations by the Geological Survey of Canada, Part 1, *Geol. Surv. Can. Paper* **61-17**, p. 15, 16, 20.

Magaritz, M., and Taylor, H.P. Jr. 1976a. Isotope evidence for meteoric-hydrothermal alteration of plutonic igneous rocks in the Yakutat Bay and Skagway areas, Alaska. *Earth Planet. Sci. Lett.* **30**, p. 179-190.

Magaritz, M., and Taylor, H.P. Jr. 1976b. $^{18}O/^{16}O$ and D/H studies along a 500 km traverse across the Coast Range batholith and its country rocks, central British Columbia. *Can. Jour. Earth Sci.* **13**, p. 1514-1536.

Magaritz, M., and Taylor, H.P. Jr. 1976c. Oxygen, hydrogen and carbon isotope studies of the Franciscan formation, Coast Ranges, California. *Geochim. Cosmochim. Acta* **40**, p. 215-234.

Magaritz, M., Whitford, D.J., and James, D.E. 1978.Oxygen isotopes and the origin of high $^{87}Sr/^{86}Sr$ andesites. *Earth Planet. Sci. Lett.* **40**, p. 220-230.

Masi, U., O'Neil, J.R., and Kistler, R.W. 1981. Stable isotope systematics in Mesozoic granites of central and northern California and southwestern Oregon, *Contrib. Mineral. Petrol.* **76**, p. 116-126.

Mason, R. 1978. Petrology of the metamorphic rocks. George

Allen & Unwin, London, 254 p.

Matsuhisa, Y. 1979. Oxygen isotopic compositions of volcanic rocks from East Japan Island Arcs and their bearing on petrogenesis. *Jour. Volc. Geotherm. Res.* **5**, p.271-296.

Matsuhisa, Y., Goldsmith, J.R. and Clayton, R.N. 1979. Oxygen isotopic fractionation in the system quartz-albite-anorthite-water. *Geochim. Cosmochim. Acta* **43**, p. 1131-1140.

Matsuhisa, Y., Matsubaya, O., and Sakai, H. 1973. Oxygen isotope variations in magmatic differentiation processes of volcanic rocks in Japan. *Contrib. Mineral. Petrol.* **39**, p. 277-288

McCrea, J.M. 1950. The isotopic chemistry of carbonates and a paleotemperature scale. *Jour. Chem. Phys.* **18**, p. 849-857.

McCulloch, M.T., Gregory, R.T., Wasserburg, G.J., and Taylor, H.P. Jr. 1980. A neodymium, strontium, and oxygen isotopic study of the Cretaceous Samail ophiolite and implications for the petrogenesis and seawater-hydrothermal alteration of oceanic crust. *Earth. Planet. Sci. Lett.* **46**, p. 201-211.

McDowell, S.D., and Elders, W.A. 1980. Authigenic layer silicate minerals in borehole Elmore 1, Salton Sea geothermal field, California, USA. *Contrib. Mineral. Petrol.* **74**, p. 293-310.

McNutt, R.H., Crocket, J.H., Clark, A.H., Caelles, J.C., Farrar, E., Haynes, S.J., and Zentilli, M. 1975. Initial $^{87}Sr/^{86}Sr$ ratios of plutonic and volcanic rocks of the Central Andes between latitudes 26° and 29° south. *Earth Planet. Sci. Lett.* **27**, p. 305-313.

Mérigoux, H. 1968. Etude de la mobilité de l'oxygène dans les feldspaths alcalins. Bull. *Soc. Fr. Minér. Crystallogr.* **91**, p. 51-64.

Merrill, R.B., Robertson, J.K., and Wyllie, P.J. 1970. Melting reactions in the system $NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2-H_2O to 20 kilobars compared with results for other feldspar-quartz-H_2O and rock-H_2O systems. *Jour. Geol.* 78, p.558-569.

Monger, J.W.H. 1975. Upper Paleozoic rocks of the Atlin Terrane, northwest British Columbia and south central Yukon. *Geol. Surv. Can. Paper* **74-47**, 63p.

Monger, J.W.H., Price, R.A., Tempelman-Kluit, D.J.

1982.Tectonic accretion and the origin of two major metamorphic and plutonic welts in the Canadian Cordillera. *Geology* **10**, p. 70-75.

Morrison, G.W. 1981. Setting and origin of Skarn deposits in the Whitehorse copper belt, Yukon. Unpub. Ph.D. thesis, Univ. of Western Ontario, London, Ontario, 306 p.

Morrison, G.W., Godwin, C.I. and Armstrong, R.L. 1979. Interpretation of isotopic ages and $^{87}Sr/^{86}Sr$ initial ratios for plutonic rocks in the Whitehorse map area, Yukon. *Can. Jour. Earth Sci.* **16**, p. 1988-1997.

Muehlenbachs, K., Anderson, A.T., Jr., and Sigvaldason, G.E. 1974. Low-O^{18} basalts from Iceland. *Geochim. Cosmochim. Acta* **38**, p. 577-588.

Muehlenbachs, K., and Byerly, G. 1982. ^{18}O enrichment of silicic magmas caused by crystal fractionation at the Galapagos Spreading Center. *Contrib. Min. Petrol.* **79**, p. 76-79.

Muehlenbachs, K., and Clayton, R.N. 1972. Oxygen isotope studies of fresh and weathered submarine basalt. *Can. Jour. Earth Sci.* **9**, p. 172-184.

Muehlenbachs, K., and Clayton, R.N. 1976. Oxygen isotope composition of the oceanic crust and its bearing on seawater. *Jour. Geophys. Res.* **81**, p. 4365-4369.

Nagy, K.L., and Parmentier, E.M. 1982. Oxygen isotopic exchange at an intrusive contact. *Earth Planet. Sci. Lett.* **59**, p. 1-10.

Nguyen, K.K., Sinclair, A.J. and Libby, W.G. 1968. Age of the northern part of the Nelson Batholith. *Can. Jour. Earth Sci.* **5**, p. 955-957.

Northcote, K. E. 1969. Geology and geochronology of the Guichon Creek Batholith. *B.C. Dept. of Mines & Petrol. Resources Bull.* **56**, 73p.

Norton, D., and Taylor, H.P. Jr. 1979. Quantitative simulation of the hydrothermal systems of crystallizing magmas on the basis of transport theory and oxygen isotope data: An analysis of the Skaegaard intrusion: *Jour. Petrol.* **20**, p. 421-486.

O'Neil J.R., Adami L.H., and Epstein S. 1975. Revised value for the ^{18}O fractionation between CO_2 and H_2O at 25°C. *Jour. Res. U. S. G. S.* **3** , p. 623-624.

O'Neil, J.R., and Chappell, B.W. 1977. Oxygen and hydrogen isotope relations in the Berridale batholith. *Jour. Geol. Soc. Lond.* **133**, p. 559-571.

O'Neil J.R., Epstein S. 1966a. A method for oxygen isotope analysis of milligram quantities of water and some of its applications. *Jour. Geophys. Res.* 71, p. 4955-4961

O'Neil J.R., and Epstein, S. 1966b. Oxygen isotope fractionation in the system dolomite-calcite-carbon dioxide. *Science* **152**, p. 198-201

O'Neil, J.R., Shaw, S.E., and Flood, R.H. 1977. Oxygen and hydrogen isotope compositions as indicators of granite genesis in the New England batholith, Australia. *Contrib. Mineral. Petrol.* **62**, p. 313-328.

O'Neil, J.R., and Taylor, H.P. Jr. 1967. The oxygen isotope and cation exchange chemistry of feldspars. *Am. Mineral.* **52**, p. 1414-1437.

Onuma N., Clayton R.N., and Mayeda T.K. 1972. Oxygen isotope temperatures of "equilibrated" ordinary chondrites. *Geochim. Cosmochim. Acta* **36**, p. 157-168.

Oxburgh, E.R., and Turcotte, D.L. 1970. Thermal structure of island arcs. *Geol. Soc. Am. Bull.* **81**, p. 1665-1688.

Perry, E.C., Ahmad, S.N., and Swulius, T.M. 1978. The oxygen isotope composition of 3800 M.Y. old metamorphosed chert and iron formation from Isukasia, West Greenland. *Jour. Geol.* **86**, p. 223-239.

Peto, P. 1974. Plutonic evolution of the Canadian Cordillera. *Geol. Soc. Am. Bull.* **85**, p. 1269-1276.

Piwinskii, A.J., and Wyllie, P.J. 1970. Experimental studies of igneous rock series: Felsic body suite from Needle Point pluton, Wallowa Batholith, Oregon, *Jour. Geol.* **78**, p. 52-76.

Preto, V.A., Osatenko, M.J., McMillan, W.J., and Armstrong, R.L. 1979. Isotopic dates and strontium isotopic ratios for plutonic and volcanic rocks in the Quesnel Trough and Nicola Belt south-central British Columbia. *Can. Jour. Earth Sci.* **16**, p. 1658-1672.

Reynolds, T.J. 1980. Variations in hydrothermal fluid characteristics through time at the Santa Rita porphyry copper deposits, New Mexico. Unpub. M.Sc. thesis, Univ. Arizona, 52p.

Roddick, J.A., Gabrielse, H., Woodsworth, G.J., Reesor,

J.E., and Hutchison, W.W. 1975. Plutonic evolution of the Canadian Cordillera: Discussion. *Geol. Soc. Am. Bull.* **86**, p. 1610-1611.

Roddick, J.A., and Hutchison, W.W. 1974. Setting of the Coast Plutonic Complex, British Columbia. *Pacific Geol.* **8**, p. 91-108.

Savin, S.M. 1980. Oxygen and hydrogen isotope effects in low-temperature mineral-water interactions: *in* P. Fritz and J.Ch. Fontes, eds., Handbook of environmental isotope geochemistry, volume 1: The terrestrial environment, A, p. 283-327.

Savin, S.M., and Epstein, S. 1970a. The oxygen and hydrogen isotope geochemistry of ocean sediments and shales. *Geochim. Cosmochim. Acta* **34**, p. 43-63.

Savin, S.M., and Epstein, S. 1970b. The oxygen isotopic compositions of coarse grained sedimentary rocks and minerals. *Geochim. Cosmochim. Acta* **34**, p. 323-329.

Sheppard, S.M.F. 1977. The Cornubian batholith, S W England: D/H and $^{18}O/^{16}O$ studies of kaolinite and other alteration minerals. *Jour. Geol. Soc. Lond.* **133**, p. 573-591.

Sheppard, S.M.F., Nielsen, R.L., and Taylor, H.P. Jr. 1969. Oxygen and hydrogen isotope ratios of clay minerals from porphyry copper deposits. *Econ. Geol.* **64**, p. 755-777.

Sheppard, S.M.F., Nielsen, R.L., and Taylor, H.P. Jr. 1971. Oxygen and hydrogen isotope ratios in minerals from porphyry copper deposits. *Econ. Geol.* **66**, p. 515-542.

Shieh, Y.N., and Schwarcz, H.P. 1974. Oxygen isotope studies of granite and migmatite, Grenville province of Ontario, Canada. *Geochim. Cosmochim. Acta.* **38**, p. 21-45.

Shieh, Y.N., and Taylor, H.P. Jr. 1969a. Oxygen and hydrogen isotope studies of contact metamorphism in the Santa Rosa range, Nevada and other areas. *Contr. Miner. Petrol.* **20**, p. 306-356.

Shieh, Y.N., and Taylor, H.P. Jr. 1969b. Oxygen and carbon isotope studies of contact metamorphism of carbonate rocks. *Jour. Petrol.* **10**, p. 307-331.

Smith, B., and Giletti, B.J. 1982. High-temperature diffusion-controlled O-exchange between quartz, feldspar and H_2O in Skye porphyritic felsites.(Abs.)

EOS Trans. Am. Geophys. Union **63** No 65, p. 1145.

Staugidel, H., Muehlenbachs, K., Richardson, S.H., Hart, S.R. 1981. Agents of low temperature ocean crust alteration. *Contrib. Mineral. Petrol.* 77, p.150-157.

Stern, C.R., W. L. Huang, and Wyllie, P.J. 1975. Basalt-andesite-rhyolite-H_2O: Crystallization intervals with excess H_2O and H_2O-undersaturated liquidus surfaces to 35 kilobars, with implications for magma genesis, *Earth Planet. Sci. Lett.* **28**, p. 189-196.

Taylor, B.E. and O'Neil, J.R. 1977. Stable isotope studies of metasomatic Ca-Fe-Al-Si skarns and associated metamorphic and igneous rocks, Osgood Mountains, Nevada. *Contrib. Mineral. Petrol.* **63**, p. 1-49.

Taylor, H.P. Jr. 1968. The oxygen isotope geochemistry of igneous rocks. *Contrib. Mineral. Petrol.* **19**, p. 1-71.

Taylor, H.P. Jr. 1971. Oxygen isotope evidence for large-scale interaction between meteoric ground waters and Tertiary granodiorite intrusions, western Cascade Range, Oregon. *Jour. Geophys. Res.* **76**, p. 7855-7874.

Taylor, H.P. Jr. 1974a. The application of oxygen and hydrogen isotope studies to problems of hydrothermal alteration and ore deposition. *Econ. Geol.* **69**, p. 843-883.

Taylor, H.P. Jr. 1974b. Oxygen and hydrogen isotope evidence for large-scale circulation and interaction between groundwaters and igneous intrusions, with particular reference to the San Juan volcanic field, Colorado. *in* A.W. Hofmann, B.J. Giletti, H.S. Yoder, Jr., and R.A. Yund, eds., Geochemical transport and kinetics. Carnegie Inst. Washington Publ. 634, p. 299-324.

Taylor, H.P. Jr. 1977. Water/rock interactions and the origin of H_2O in granitic batholiths. *Jour. Geol. Soc. London* **133**, p. 509-558.

Taylor, H.P. Jr. 1978. Oxygen and hydrogen isotope studies of plutonic granitic rocks. *Earth Planet. Sci. Lett.* **38**, p. 177-210.

Taylor, H.P. Jr. 1980. The effects of assimilation of country rocks by magmas on the $^{18}O/^{16}O$ and $^{87}Sr/^{86}Sr$ systematics in igneous rocks. *Earth Planet Sci. Lett.* **47**, p. 245-254.

Taylor, H.P. Jr., and Forester, R.W. 1979. An oxygen and hydrogen isotope study of the Skaergaard intrusion and its country rocks: A description of a 55 Ma old fossil hydrothermal system. *Jour. Petrol.* **20**, p. 355-419.

Taylor, H.P. Jr., and Magaritz, M. 1978. Oxygen and hydrogen isotope studies of the Cordilleran batholiths of western North America: *in* Robinson, B.W., ed., Stable Isotopes in the Earth Sciences, *DSIR Bull.* **220**, p. 151-173, New Zealand Dept. of Scientific and Ind. Res., Wellington, New Zealand.

Taylor, H.P. Jr., Silver, L.T. 1978. Oxygen isotope relationships in plutonic rocks of the Peninsular Ranges batholith, Southern and Baja California. *In* Short papers of the Fourth International Conference, Geochronology, Cosmochronology, Isotope Geology. p. 423-426.

Tempelman-Kluit, D.J. 1976. The Yukon Crystalline Terrane: enigma in the Canadian Cordillera. *Geol. Soc. Am. Bull.* **87**, p. 1343-1357.

Tempelman-Kluit, D.J. 1979. Transported cataclasite, ophiolite and granodiorite in Yukon: Evidence of arc-continent collision: *Geol. Surv. Can. Paper* **79-14**, 27p.

Tempelman-Kluit, D.J., and Wanless, R.K. 1975. Potassium-argon age determinations of metamorphic and plutonic rocks in the Yukon Crystalline Terrane. *Can. Jour. Earth Sci.* **12**, p. 1895-1909.

Tenney, D. 1981. The Whitehorse copper belt. Mining Exploration and Geology (1967-1980). *D.I.A.N.D. EGS Bull.* **1**, publ. # QS-Y003-000-EE-AI, Ottawa, 29p.

Titley, S.R., and Beane, R.E. 1981. Porphyry copper deposits, Part I: Geological settings, Petrology and tectogenesis. *Econ. Geol. 75th Anniversary Volume*, p. 214-234.

Turi, B., and Taylor, H.P. Jr. 1971. An oxygen and hydrogen study of a granodiorite pluton from the Southern California batholith: *Geochim. Cosmochim. Acta* **35**, p. 383-406.

Wanless, R.K., Loveridge, W.D., and Mursky, G. 1968. A geochronological study of White Creek Batholith, Southeastern British Columbia. Can. Jour. Earth Sci. **5**, p.375-386.

Wanless, R.K., and Lowdon, J.A. 1963. K-Ar measurements on

mineral pairs, *In* Age Determination by the Geological Survey of Canada, Part 1, *Geol. Surv. Can. Paper* **62-17**, p. 29.

Watson, P.H., Godwin, C.I., and Armstrong, R.L. 1981. Geology, mineralization and K-Ar and Rb-Sr isotopic study of the Ram zinc-lead-silver property Yukon Plateau, Southwest Yukon Territory. *in* Yukon Geology and Exploration 1979-1980, *D.I.A.N.D. EGS publ. #* QS 8278-000-E1-A1, Ottawa, p. 123-127.

Wheeler, J.O. 1961. Whitehorse map-area, Yukon Territory. *Geol. Surv. Can. Mem.* **312**, 156p.

White, W.H., Sinclair, A.J., Harakal, J.E., and Dawson, K.M. 1970. Potassium-argon ages of Topley intrusions near Endako, British Columbia. *Can. Jour. Earth Sci.* 7, p. 1172-1178.

Whitney, J.A. 1975. The effects of pressure temperature and XH_2O on phase assemblages in four synthetic rock compositions, *Jour. Geol.* **83**, p. 1-31.

Wilkinson, W.H. Jr. 1981. The distribution of alteration and mineralization assemblages of the Mineral Park mine, Mohave County, Arizona. Unpub. Ph.D. thesis, Univ. of Arizona, Tucson, 101p.

Wilson, A.F., and Green, D.C. 1970. The use of oxygen isotopes for geothermometry of Proterozoic and Archaean granulites. *Spec. Publ. Geol. Soc. Aust.* **3**, p. 389-400.

Winkler, H.G.F. 1979. Petrogenesis of metamorphic rocks. 5th ed. Springer-Verlag, New York. 348p.

Wyllie, P.J., W. L. Huang, C. R. Stern, and Maaloe, S. 1976. Granitic magmas: Possible and impossible sources, water contents, and crystallization sequences. *Can. Jour. Earth Sci.* **13**, p. 1007-1019.

Wyllie, P.J., and Tuttle, O.F. 1961. Hydrothermal melting of shales, *Geol. Mag.* **98**, p. 56-66.

Yund, R.A., Smith B.M., and Tullis, J. 1981. Dislocation-assisted diffusion of oxygen in albite. *Phys. Chem. Min.* 7, p. 185-189.

APPENDIX 1

Petrographic summary of thin sections

Sample #	QTZ %	KFP %	PFP %	BIO %	HBL %	CHL %	SAU %	OTHER %	Gr.Size (Fp)mm
WHA 1	30	55"	15	1	--	--	--	"40 kao	4.0
WHA 2a	5	5	55	10*	30	*1		1 opa	1.4
WHA 2c	15	1	60#	20*	5	*1	#1	tr.opa	1.5
WHA 3a	25	30"	40#	5*	1	*4	#15	"5 kao	2.7
WHA 3b	35	40"	25#	1	--	--	#5	"25 kao	0.9
WHA 3c	35	10"	50#	2*	1	*1	#30	"1 kao	2.8
WHA 4	25	30"	41	3*	3	*1	--	"4 kao	2.4
WHA 5	5	30"	40#	1	--	*6	#20	25*cpx	2.0
WHA 6a	10	35"	45"	--	3*	*1	--	"25 kao	2.2
WHA 6c	5	75#		--	*3	*2	#70	10*cpx	1.3
WHA 6E	15	85#		--	--	*1	#10	#50 kao	0.9
WHA 7	25	20	50	2*	1	*tr.	tr.	1 opa	2.3
WHA 8a	25	25	40	8	2	1	--	tr.epi	3.5
WHA 8c	2	3	25#	--	45	--	#3	25 cpx	0.5
WHA 9	17	32	42#	4*	2	*1	#4	tr.opa	3.7
WHA 10	6	4	50	15	*15	*3	--	25*cpx	2.0
WHA 11	15	25	55#	--	5*	*5	#28	1 opa	2.5
WHB 1	12	10	55#	5	15	tr.	#5	tr.opa	2.0
WHB 2	20	10	55#	2*	10	*1	#3	1 opa	1.5
WHB 3	20	12	50#	5*	10	*3	#6	2 act	1.4
WHB 4	35	20	40#	2*	3	*1	#20	tr.epi	1.8

APPENDIX 1 (continued)

Sample #	QTZ %	KFP %	PFP %	BIO %	HBL %	CHL %	SAU %	OTHER %	Gr.Size (Fp)mm
WHB 5	17	10	50#	5*	15	*2	#2	tr.opa	1.8
LW77A6C	20	5	50#	7	6	--	#1	10 epi	3.0
LW77B9C	30	15	45#	--	5	*5	#8	--	6.0
LW77A37B	25	30	35#	4	6	--	#8	--	3.5
LW77A58B	20	7	55#	8*	5	*1	#3	1 shp	2.5
C 1	28	5	50#	15*	--	*1	#1	1 epi	4.5
C 4	12	4	60#	7*	16	*1	#1	tr.opa	1.5
C 6	10	--	65#	15	8*	--	#tr.	*2 epi	2.5
K 1	15	--	60#	10*	15	*2	#1	#1 epi	4.5
K 3	33	4	43#	7*	12"	"*1	#3	"1 epi	4.0
P 1	25	12	55	5*	3"	*2	1	"1 epi	4.0
P 2	25	9	55#	10*	3"	*"1	--	"#2 epi	3.0
N 1	30	70"		1	--	--	--	"20 kao	3.0
R 2	20	15"	55"	4*	7	*tr.	--	"10 kao	4.0
R 3	30	15"	45"#	4*	--	*2	#2	"2 kao	1.5
JKqm-1	12	18	45#	8*	17"	*1	#8	"1 epi	3.5
JKqm-2	2	5	65#	15*	10	*6	#55	10 cpx	3.0
JKdi-1	--	--	50#	--	10	*15	#10	25*hyp	1.0
Mgd-1	8	1	75	6	*3	--	--	6* opx	3.0
Mgd-3	15	2	70#	6	5	--	#2	tr.epi	8.0
Mgd-4	25	60	12#	2*	--	*1	#1	1 opa	2.5
Mgd-5	12	25"	54#	--	--	7	#30	"10 kao	2.5

APPENDIX 1 (continued)

Sample #	QTZ %	KFP %	PFP %	BIO %	HBL %	CHL %	SAU %	OTHER %	Gr.Size (Fp)mm
Mgdn-6	15	45	40#	2*	--	*1	#20	tr.	5.0
Mgdn-7	10	5	60#	--	20*	*12	#60	*5 epi	2.5
Kqm-1a	18	45"	35#	5	--	--	#1	"4 kao	3.5
Kqm-2	20	35"	40#	7	--	--	#2	"4 kao	1.5
Kqm-3	30	50	15#	5*	--	*2	#2	tr.opa	4.5
Kqm-4	23	57"	15#	3	--	--	#3	"19 kao	3.0
Kqm-5	25	55	15	4	--	--	--	tr.opa	2.5
Kqm-6	17	55"	20#	5*	--	*4	#8	"5 kao	8.0
Kqm-7	30	50	15#	5	--	--	#7	1 mus	5.0
Kgal-1	35	50"	10#	3	--	--	#9	"8 kao	7.0
Kgal-2	na	na	na	8	3	--	--	"5 kao	18
Kgal-3	na	na	na	5	--	--	#3	"10 kao	6.5
Kgal-4	15	78"	2	5	--	--	--	"6 kao	11
KTgd-1	10	10	65#	5*	10"	*3	#10	"2 epi	2.1
KTgd-3	5	2	70#	1	20	--	#5	tr.opa	2.2
Tqm-1	10	5	50#	4	30*	*25	#30	#4 cal	1.5
lTg-1	15	50"	30	6*	--	1*	tr.	"5 kao	2.0
lTg-2	20	55	20	3	3	--	--	tr.opa	5.0
lTg-3	25	50	20#	3*	3	*1	#8	tr.opa	4.5
43-D	25	5	65#	4*	--	*3	#50	--	
1075	10	35"	55#	3*	--	*1	#40	"10 kao	1.0
KTP	20	45"	35#	2	--	--	#25	"20 kao	3.0

APPENDIX 1 (continued)

Sample #	Gr.Size
WHA 6b	1.0
WHA 6d	0.7
KTgd-4	1.8

The symbols #, *, " indicate the amounts and types of of alteration present in the primary phases of the rock. The position of the symbols indicate whether the mineral is primary and altered (on the right hand side of the % number) or if it is a product of the alteration (left side of % number). Mineral abbreviations are as before, except for the following: epi = epidote, opa = opaques, kao = clays, cpx = clinopyroxene, opx = orthopyroxene, hyp = hypersthene, fp = feldspar, sph = sphene, cal = calcite, mus = muscovite, and SAU = saussurite. The Gr. Size column shows the average feldspar grain size for each sample as measured from thin sections and hand specimens.

APPENDIX 2

Locations and descriptions of plutonic samples

Sample #	Longtitude deg min	Latitude deg min	Description
WHA 1	135 53.3	60 50.6	CG, Leu, Bio Syenogranite
WHA 2a	134 47.5	60 48.7	MG, Hbl Bio Qtz Diorite
WHA 2b	134 47.6	60 48.4	FG, Hbl Bio Qtz Diorite
WHA 2c	134 47.6	60 48.2	MG, Bio Hbl Tonalite
WHA 3a	134 47.5	60 48.0	CMG, Bio Hbl Qtz Monzonite
WHA 3b	134 49.7	60 48.6	FG, Leu, Bio Monzogranite
WHA 3c	134 49.5	60 50.2	CMG, Leu,Bio Granite Porph
WHA 4	135 22.2	60 41.4	CG, Bio Hbl Porph Granite
WHA 5	135 10.6	60 38.8	MG, Augite Bio Monzonite
WHA6a-f	135 08.3	60 37.2	MFG, Hbl Qtz Monzo Granoph
WHA 7	134 44.7	60 30.3	MCG, Bio Hbl Granodiorite
WHA 8a	135 24.5	60 05.5	MCG, Bio Hbl Granodiorite
WHA 8b	135 23.7	60 08.3	CG, Bio Qtz Monzonite
WHA 8c	135 24.1	60 07.1	MG, Hbl Aug Porph Diorite
WHA 9	134 42.7	60 09.0	CG, Bio Hbl Qtz Monzonite
WHA 10	135 54.4	60 16.9	CMG,Bio Augite Qtz Diorite
WHA 11	135 26.2	60 43.7	MG, Hbl Qtz Monzodiorite
WHB 1	135 09.7	60 43.7	MCG, Hbl Bio Qtz Monzodior
WHB 2	135 06.1	60 39.8	MG, Hbl Bio Granodiorite
WHB 3	134 57.9	60 34.8	MG, Hbl Bio Granodiorite
WHB 4	134 57.4	60 34.6	MG, Hbl Bio Granodio Porph

APPENDIX 2 (continued)

Sample #	Longtitude deg min		Latitude deg min		Description
WHB 5	134	54.7	60	34.9	MG, Hbl Bio Granodiorite
KTgd-1	135	08.5	60	44.8	CMG, Hbl Bio Qtz Diorite
KTgd-2	135	09.2	60	44.9C	MCG, Hbl Bio Qtz Diorite
KTgd-4	135	11.0	60	43.8	MCG, Hbl Bio Granodiorite
LW77A6C	134	18.3	59	26.0	MG, Hbl Bio Granodiorite
LW77B6D	134	17	59	25.8	CG,Hbl Bio Porph Granodior
LW77B9C	133	59.1	59	09.1	CG,Hbl Bio Porph Granodior
LW77A37B	134	01.4	59	12.8	CG, Hbl Bio Porph Granite
LW77A58B	134	11.3	59	13.3	CG, Hbl Bio Tonalite
T741081	134	05.5	59	15.8	CG, Hbl Bio Granodiorite
T742071	134	22.7	59	24.8	CG, Bio Hbl Granodiorite
T742081	134	17.1	59	25.5	CG, Hbl Bio Qtz Monzonite
T742281	134	31.9	59	33.0	CG, Hornblende Tonalite
T751014	134	42.5	59	55.3	CG, Bio Hbl Qtz Monzonite
T751022	134	46.3	59	51.9	CG, Bio Hbl Qtz Monzonite
T751302	134	07.0	59	17.9	CG,Hbl Bio Porph Granodior
T753102	134	20.0	59	48.9	CG,Bio Hbl Porph Granodior
T753232	134	26.2	59	25.3	CG, Hornblende Qtz Diorite
T753233	134	26.2	59	25.3	CG, Biotite Granite
T754131	134	24.5	59	47.4	CG,Bio Hbl Porph Granodior
C 1	138	23.4	62	41.3	CG, Bio Tonalite
C 4	138	53.0	62	46.4	FMG, Hbl Bio Qtz-diorite

APPENDIX 2 (continued)

Sample #	Longtitude deg min		Latitude deg min		Description
C 5	138	52.8	62	46.9	CG, Hbl Bio Granodiorite
C 6	138	53.0	62	47.6	MG, Hbl Bio Qtz-diorite
K 1	137	12.4	61	42.0	CG, Hbl Bio Tonalite
K 3	138	32.6	62	17.3	CMG,Hbl Bio Porph Tonalite
P 1	136	40.0	61	51.6	CG,Hbl Bio Porph Granodior
P 2	163	40.4	61	46.4	CG,Hbl Bio Porph Granodior
N 1	137	37.9	61	30.7	MG, Leuco, Bio Alaskite
R 2	137	45.0	61	20.9	CG, Hbl Bio Granodiorite
R 3	136	59.7	61	04.9	MG, Leu, Bio Granodiorite
RW371	135	45.0	60	11.6	CG, Hbl Bio Granodiorite
JKqm-1	133	47.9	59	48.8	CMG, Hbl Bio Qtz Monzodior
JKqm-2	133	43.8	59	39.5	MG, Bio Hbl Monzonite
JKdi-1	133	27.6	60	28.1	MG, Hypersthene Epidiorite
Mgd-1	137	00.0	60	31.1	MCG, Bio Hbl Qtz Diorite
Mgd-3	136	53.1	60	02.3	CG, Bio Hbl Qtz Diorite
Mgd-4	136	09.4	60	33.8	MG, Bio SyenoGranite
Mgd-5	134	57.5	60	03.8	CMG, Chl Qtz Monzodiorite
Mgdn-6	135	12.1	62	07.4	CG, Bio Porph Qtz Monzo
Mgdn-7	133	47.6	59	56.1	MG, HBl Pyrox Epidiorite
Kqm-1a	134	02.1	62	10.9	MCG, Leu,Bio Qtz Monzonite
Kqm-2	134	00.2	62	10.7	MG, Bio Qtz Monzonite
Kqm-3	133	08.7	60	33.7	CG, Bio Qtz Monzonite

APPENDIX 2 (continued)

Sample #	Longtitude deg min	Latitude deg min	Description
Kqm-4	133 05.5	60 58.1	MG, Bio SyenoGranite
Kqm-5	132 36.8	60 10.8	MG, Bio SyenoGranite
Kqm-6	130 40.8	60 05.7	CG, Bio Qtz Monzonite
Kqm-7	130 27.5	60 03.2	MG, Bio Musc Porph Granite
Kgal-1	133 20.1	59 40.1	CG, Bio Granite Porphyry
Kgal-2	132 57.1	59 36.6	CG, Bio Hbl Alaskite
Kgal-3	132 56.2	59 40.0	CG, Biotite Alaskite
Kgal-4	133 11.4	59 36.5	CG, Biotite Alaskite
Tqm-1	132 26.6	59 58.9	CG, Bio Hbl Qtz Diorite
lTg-1	136 08.8	60 29.0	CG,Bio Porph Qtz Monzonite
lTg-2	136 08.8	60 29.3	MG, Hbl Bio Granite
lTg-3	133 25.6	60 27.1	CG, Hbl Bio Granite
A-8-1-5	133 44.2	59 26.5	FG, Bio Hbl Tonalite
1075	129 51.0	59 20.6	MFG, Qtz Monzo Porphyry
KTP	129 52.0	62 46.6	CG, Bio Qtz Monzonite
43-D	131 52.6	62 30.7	MG, Biotite Tonalite

MG, indicates a medium grain size for the sample while F and C respectively indicate fine and coarse-grained varieties. "Porph" either means a porphyritic rock type or if it is preceded by a rock name; the type of porphyry. "Leuco" or "Leu" is the abbreviation of leucocratic. Rock type abbreviations are as follows: Granoph = Granophyre, Dior = Diorite, Monzo = Monzonite. Mineral abbreviations are described in the introduction of the text except for Musc, which represents muscovite.

APPENDIX 3

Oxygen isotope data from the Whitehorse batholith and skarns as reported in G. Morisson's dissertation

Sample #	Pluton	Early Veining	Diopside Skarn	Garnet Skarn	Late Veining
Gr21A	8.2	-12.5	-3.4	-5.7	
Pb13A	7.3	-11.4			
BC19	8.5	- 7.4	-9.8		
76-24	8.8		-5.6		
G140				3.6 qtz	
AC41-248				2.8 qtz	
WE6					-13.3 qtz

Analyses were done by Dr. R. Kerrick at the University of Western Ontario for G. Morrison. All results are bulk rock or bulk vein unless otherwise specified. All values shown are in ‰ relative to SMOW. Sample locations are described in Morrison(1981, appendix 7).

APPENDIX 4

Locations of collected country rock samples

Sample #	Longtitude deg min	Latitude deg min
HC-1	134 09.1	62 10.2
HC-2	134 00.2	62 10.7
HC-3	136 58.5	61 08.6
HCsn-1	136 59.6	61 09.3
HPm-1	134 02.1	62 10.9
PPsn-1	138 52.8	62 47.7
lCc-1	130 24.2	60 05.4
CPsn-1	134 47.0	62 12.3
CPsn-3	133 02.1	60 57.4
CPsn-4	132 22.6	59 58.5
CPsn-5	132 16.2	60 02.3
CPv-1	134 41.0	60 08.1
CPv-2	134 41.5	60 04.7
CPv-4	133 27.6	60 28.1
DMs-2	130 50.5	60 04.9
Tv-1	135 33.8	62 04.9
Tv-2	132 36.8	60 10.8
uTlw-1	134 55.1	60 34.0
uTlw-3	135 11.4	60 38.9
uTlw-4	135 11.7	60 39.0
uTlv-1	134 57.9	60 04.8

APPENDIX 4 (continued)

Sample #	Longtitude deg min	Latitude deg min
uTlv-2	135 47.8	60 07.1
uTlv-3	135 47.0	60 07.7
uTlv-4	135 46.7	60 08.1
uTc-1	134 46.4	60 34.4
uTc-2	134 55.1	60 34.0
uTc-3	135 09.0	60 45.0
uTc-4	135 11.0	60 43.8
uTc-5	135 04.1	60 38.0
uTc-6	135 08.3	60 44.8
uTc-7	135 07.7	60 40.3
Jl-1	134 41.8	60 04.7
Jl-3	135 42.9	62 02.8
JKd-1	137 03.2	60 22.4
JKk-1	136 56.5	60 33.8
JKk-2	136 58.3	60 17.6
JKk-3	136 16.8	60 24.7
JKk-4	136 08.5	60 36.5
KTgd 4-3	135 11.0	60 43.8